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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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# EARTH MOTIONS IN THE VICINITY OF A FAULT SLIP

BY N. H. HECK AND FRANK NEUMANN

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## ABSTRACT

Existing lack of knowledge regarding the mechanism of fault slipping and resultant earth motions is in part met by new and important displacement data now available

for the Imperial Valley earthquake of 1940 and the Long Beach earthquake of 1933. The ground displacements were not recorded directly but obtained by mathematical treatment of seismograph acceleration records. The accuracy of the original and deduced records has been proved adequate.

In the case of the El Centro record of the Imperial Valley earthquake maximum accelerations of 0.3 g, velocity of 24 inches a second, and displacement of 15 inches are all large and indicate greater values in stronger earthquakes. The record was obtained 7 miles from the epicenter and 5 miles from the nearest point of the fault on which slipping occurred. All major wave types appear to be present, and they appear to have their genesis at a point or in a small area. The wave motions are not dominated by the direction of fault slip but are related to the direction from epicenter to recording instrument. The results show how improved records would make it possible to measure permanent displacement when it occurs. Doubt is thrown on the presence of really dominant wave periods during an earthquake in California.

## INTRODUCTION

Faults are of major importance to geologists, but little is known about the actual mechanism of their formation, especially of those buried from sight. The region in the vicinity of the fault slip is the meeting place of the geologist and the seismologist, since it is here that earthquake waves are set up. The engineer would like to know how the motion in the vicinity of the fault is divided between oscillatory and permanent displacements.

No claim to complete solution is made. However, the presentation of new data, not obtainable from ordinary seismograph records, will be of value as indicating the opening of a new field of investigation, and some preliminary conclusions can be reached. Harry Fielding Reid (1931; 1933), in studying some of the problems chiefly from theoretical considerations, expressed the view that observational data are badly needed.

It is possible that conditions at the place where the slipping occurred, possibly at a depth of 5 to 7 miles, may not be the same as at the surface, but in southern California the fault systems are so well established that it is reasonable to assume that the general directions are the same.

## DESCRIPTION OF INSTRUMENTS AND RECORDS

### GENERAL CONSIDERATIONS

The strong-motion accelerograph (Heck, 1938), as developed by the Coast and Geodetic Survey, and which produced the records to be discussed, consists of three accelerometers, two set in the horizontal plane at right angles to each other and one in the vertical. The recording is photographic and begins when the mechanical starting device is operated by the earthquake. There is always some loss of record, but there was a little more than usual in the case of the records discussed, since the recorders had to be made sufficiently insensitive that known nonseismic disturbances would not set them off. Recording stops automatically

after about one minute. Some accelerometers have an attachment which makes it possible to record the greatest accelerations likely to occur, provided the instrument survives.

Since the photographic records under discussion were only 6 inches wide, some record has been lost by the light beam going off the sheet. Many instruments now have records 12 inches wide to prevent this. These individual instruments are called accelerometers because the period of 0.1 sec. is small compared with that of earthquake waves of chief engineering interest—from ½ second to 4 seconds. In such cases the recorded acceleration varies little from the true acceleration.

In order to define the motion of the earth particle, it is also necessary to know the velocity and the displacement. So-called displacement meters with 10-second period are capable of recording displacements, with little correction, for the same range of periods just mentioned. The displacement referred to is oscillatory and not permanent. These instruments, however, are large and difficult to install and record only horizontal motion. Only 6 have been installed, whereas 45 accelerographs are in operation: 35 widely spaced in California; 4 in Nevada (3 Bureau of Reclamation in Lake Mead region); 4 in Montana; 1 in Utah; and 1 in the Panama Canal Zone. The chance that accelerations of an important earthquake will be measured is therefore great.

Since past data have been, and future data are likely to be, directly due to accelerographs, it is important to compute velocity and displacement using data taken from the accelerograms. The mathematical principle is simple, but since we are dealing with something quite different from simple harmonic motion it has required several years to perfect the integration methods especially with reference to scaling records and selecting proper axes. This is completely described by Neumann (1941).

The validity of the records and the derived curves have been tested in various ways. The accelerations have been proved sufficiently accurate by shaking-table tests, and the integration methods have been investigated and approved by competent, disinterested mathematicians. Displacement curves from both accelerographs and displacement meters of the submarine earthquake of December 20, 1940, off Cape Mendocino, California, are in good agreement. It can then be concluded that displacement curves computed from accelerograms quite accurately represent the actual ground motion during strong earthquakes.

### LONG BEACH EARTHQUAKE OF MARCH 10, 1933

The essential facts about the Los Angeles record of the Long Beach earthquake of March 10, 1933, are given in Figure 1. The three records of this shock were among the first obtained, did not have base lines,

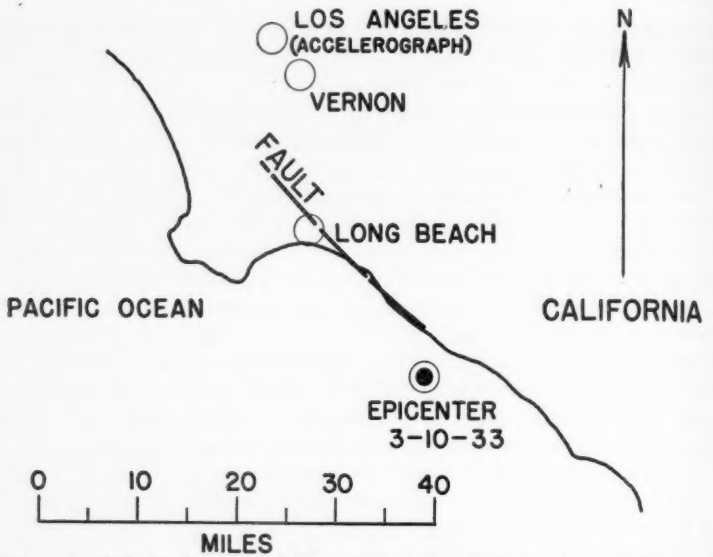



FIGURE 1.—*Relation between epicenter of Long Beach earthquake of March 10, 1933, fault, and accelerograph*

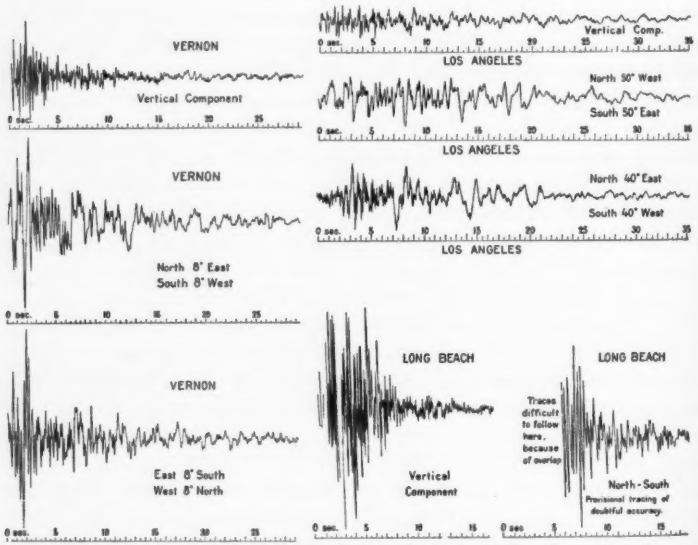



FIGURE 2.—*Tracings of accelerograph records of Long Beach earthquake*

and some were crowded and hard to disentangle. For this reason the records obtained near the epicenter at Long Beach, and in the Central Manufacturing District near Vernon, have not been used in this investigation. The Los Angeles instrument was 37 miles from the epicenter

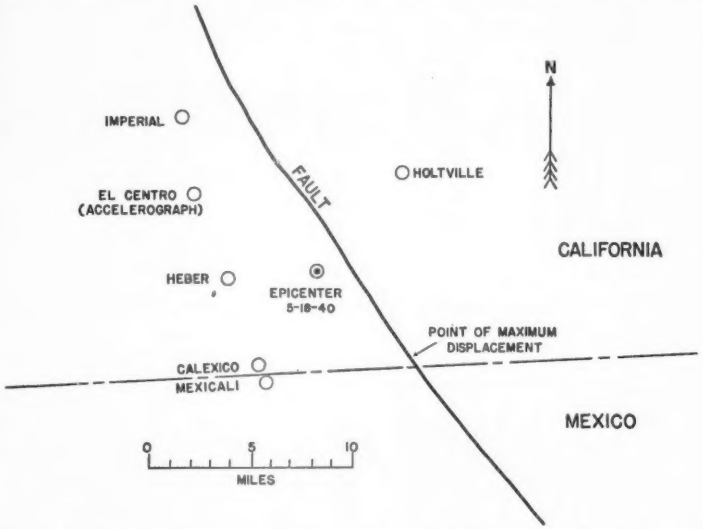



FIGURE 3.—*Relation between epicenter of Imperial Valley earthquake of May 18, 1940, fault, and accelerograph*

which, though under the sea, was accurately located by the network of seismograph stations operated in conjunction with the Seismological Laboratory at Pasadena as on a possible extension of the Inglewood fault. The direction from instrument to epicenter was 154°, the direction of the fault 135°. The accelerometers were set in directions 40°, 130°, and vertical. Tracings of the Long Beach earthquake accelerograms are shown in Figure 2.

### IMPERIAL VALLEY EARTHQUAKE OF MAY 18, 1940

The facts about the El Centro record of May 18, 1940, are given in Figure 3. Of special interest are the direction from instrument to earthquake epicenter, 128°, the direction of the fault, 140°, and the position of maximum slip at the surface, on the order of 15 feet, the southwest side moving northwestward relative to the northeast side. The true nature of this displacement, relative to the original undisturbed

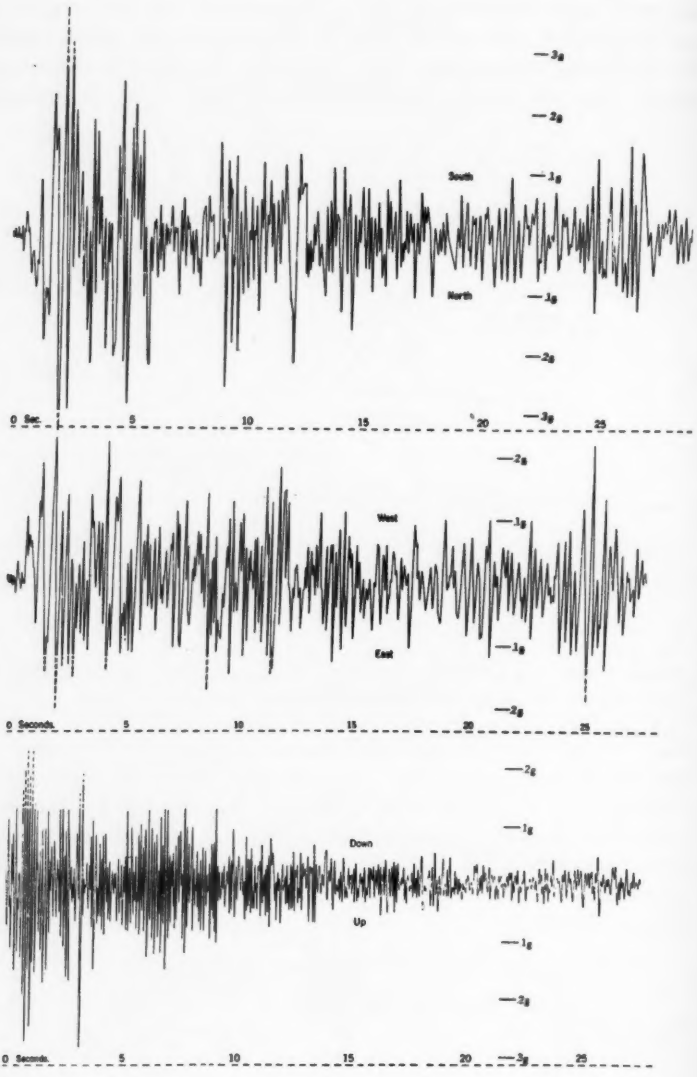



FIGURE 4.—*Tracings of acceleration records obtained at El Centro, Imperial Valley earthquake*

state, is not known. The accelerometers were in directions 0°, 90°, and vertical. This record has base lines but is only 6 inches wide.

The El Centro accelerations are shown in Figure 4. In general, they accentuate the short-period motions of 0.3 second and less and accelerations as high as 0.3 g which have so little displacement that they are insignificant on the displacement records. On the other hand, the longer waves that appear in the displacement records have so little acceleration that they cannot be seen on the acceleration records.

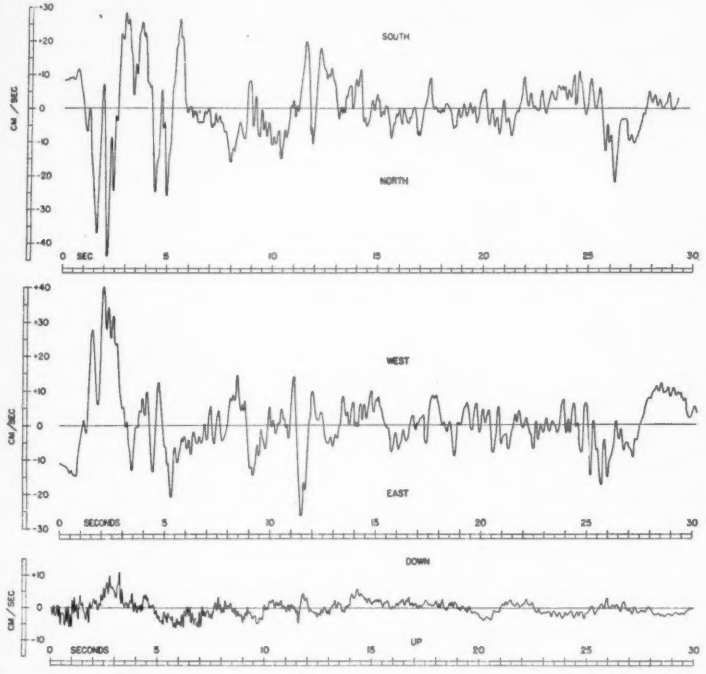



FIGURE 5.—*Velocity curves computed from El Centro accelerograph record*

## METHODS AND RESULTS OF ANALYSIS

The first essential to accurate integration is a base line from which accurate scalings can be made. This provides a satisfactory means of computing a mathematical axis. The basic principle of the double-integration process is that the axes of the acceleration and the velocity curves must be adjusted, by trial, until the second integral, or displacement curve, appears as an approximately symmetrical oscillatory

motion about a central axis. The first step is to open up the record so that ordinates can be scaled with high precision. This is done by means of an enlarging apparatus (Pl. 1), functioning somewhat as a pantograph. It accurately enlarged the El Centro record 25 times horizontally and 1½

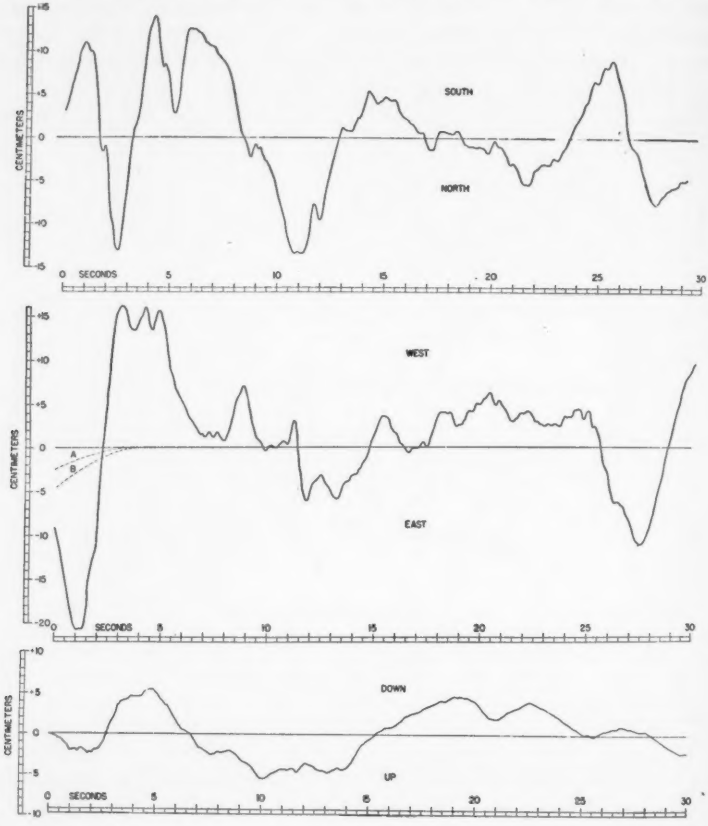
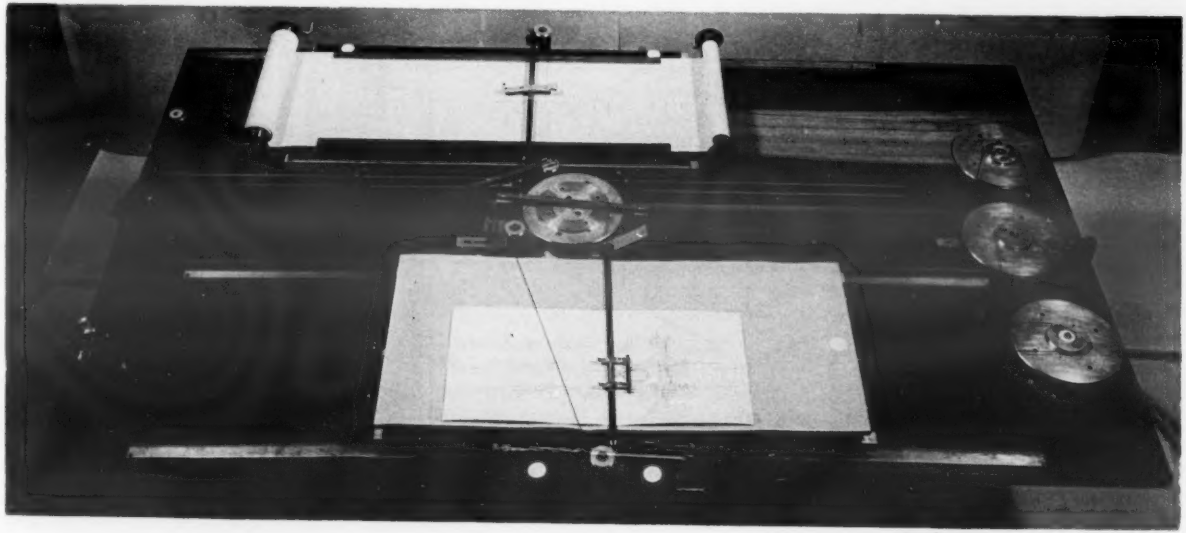



FIGURE 6.—*Displacement curves computed from El Centro accelerograph record*

times vertically. Mean ordinates were measured for each $\frac{1}{25}$ second for 30 seconds of record, making a total of 750 increments for each component. The integration computations were made on a double-register adding machine.

The results of one integration are shown in the velocity curves of Figure 5. Any point represents instantaneous velocity of the earth-



APPARATUS FOR EXPANDING ACCELERATION CURVES PREPARATORY TO DOUBLE INTEGRATION

particle motion. These curves show the intermediate periods of the earth motion most clearly and also indicate roughly the relative degree of energy of the earthquake. Maximum velocities were 18, 15½, and 5 inches per second with resultant horizontal maximum velocity of 24 inches per second.

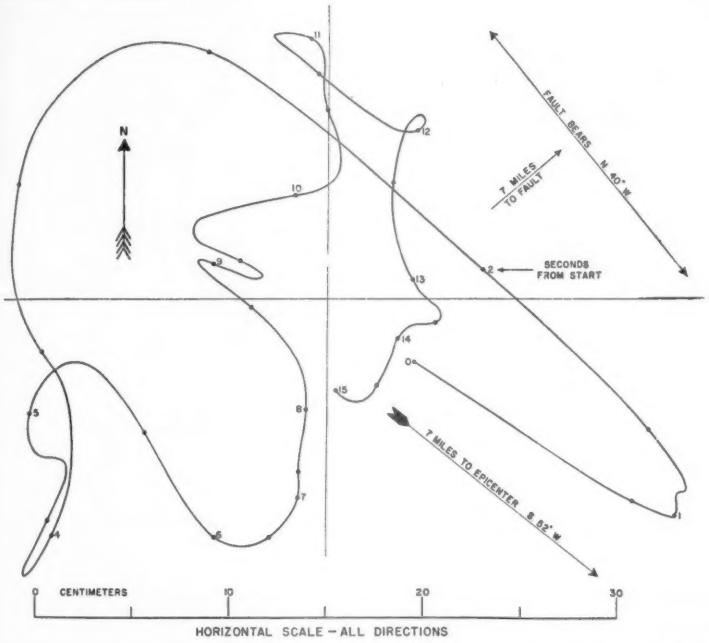



FIGURE 7.—*Resultant horizontal motion at El Centro during first 15 seconds of Imperial Valley earthquake*

Another integration produced the displacement curves shown in Figure 6. The maximum resultant displacement from position of rest was 8 inches, or a total range of motion of 15 inches. The long-period waves are clear, and the short-period waves are not evident. The maximum displacement was associated with a wave of about 3.5 seconds period. Maximum vertical displacement was about 2.2 inches, associated with a period of about 10 seconds.

The determination of the earth-particle motion at El Centro is of interest not only because of the large amplitude but because it suggests the inference of even larger amplitudes in stronger earthquakes. Further analysis of the El Centro record, and the Los Angeles record of the

Long Beach earthquake, with some justifiable speculation, permits additional conclusions to be drawn. These conclusions should be further tested by analysis of future comparable records.

Analysis of displacement is best accomplished through a vector diagram as in Figure 7, showing resultant horizontal motion, since this

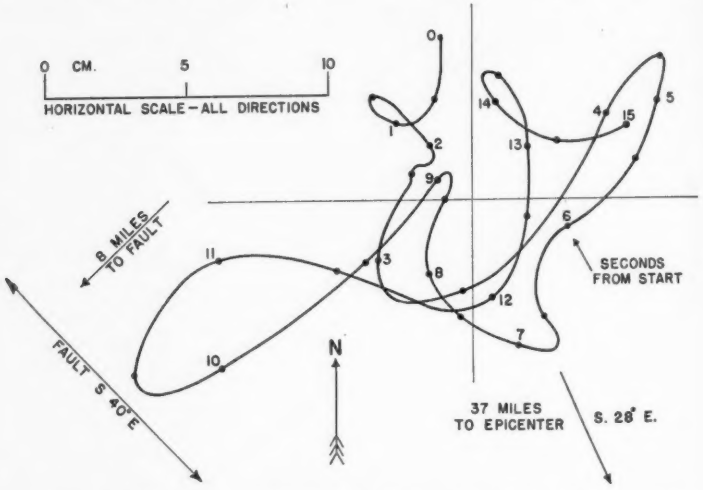



FIGURE 8.—*Resultant horizontal motion at Los Angeles during first 15 seconds of Long Beach earthquake*

eliminates the effect of directions in which accelerometers for recording horizontal motion are set. This makes it possible to compare motions with relation to direction of fault or to line from station to epicenter. The diagram includes only half, or the first 15 seconds, of the integrated record.

The elastic-rebound theory of Reid would seem to indicate that the principal earth motion is chiefly parallel to the fault, but this refers to permanent displacement. It should be understood that in a strong-motion record obtained only 7 miles from the epicenter the different types of waves arrive almost together and without time for the earlier types to lose energy before the next type arrives; also, there is uncertainty in the amount of record lost at the start. The horizontal motions on the El Centro record are apparently all of an oscillatory pattern, and the record, possibly because of its incompleteness, gives no evidence of permanent displacement at El Centro. Inspection of the record indicates that the first longitudinal motion is in the direction from

epicenter to station and roughly parallel to the fault. However, it is followed by a more complex motion which is generally at right angles to the direction from station to epicenter and therefore roughly normal

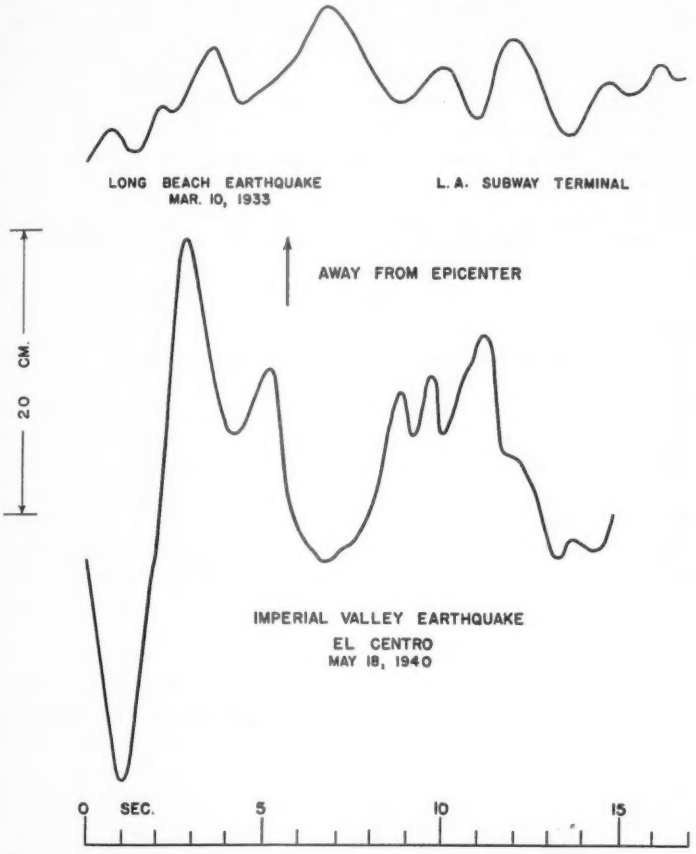



FIGURE 9.—*Longitudinal motion at Los Angeles and El Centro during the Long Beach and Imperial Valley earthquakes*

to the fault. This points to the presence, even though they cannot be fully separated, of the wave types ordinarily recorded at greater epicentral distances.

It is reasonable to assume that the record would be much more confused and that the two dominant directions of motion would not

be evident if the release of energy came from simultaneous slipping at two or more points along the fault plane. It must therefore come from a point or small area.

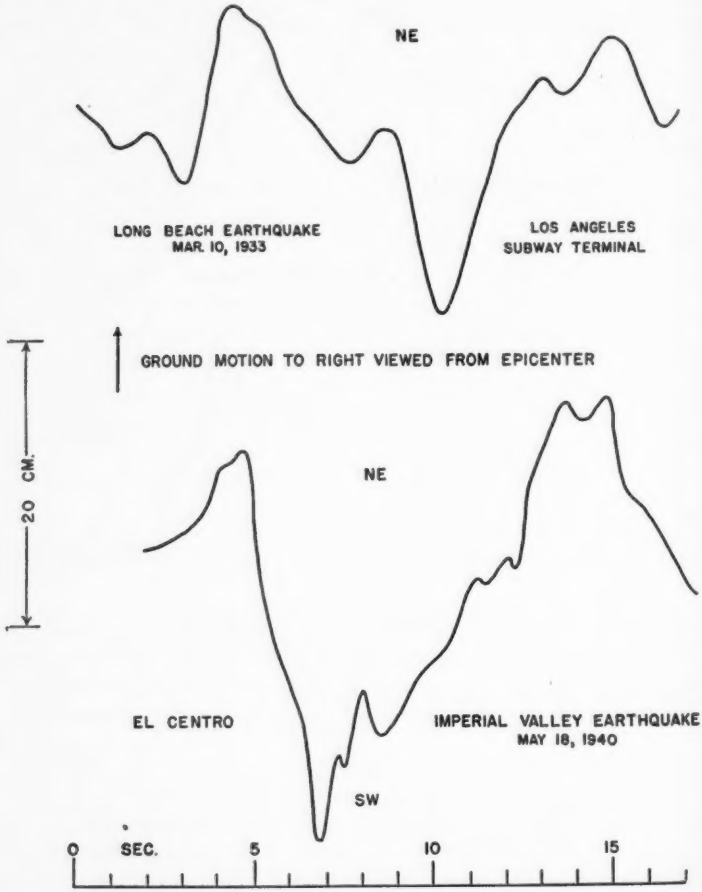



FIGURE 10.—*Transverse motion at Los Angeles and El Centro during the Long Beach and Imperial Valley earthquakes*

Though the steps are not shown, a similar diagram (Fig. 8) is obtained by the same method in the case of the Los Angeles record of the Long Beach earthquake. The distance from the epicenter is greater, but the

presence of outstanding motion at right angles to the fault is again a factor.

From the two vector diagrams, the displacements in the direction from epicenter to recording station are derived. Motion in this direction is shown in Figure 9; and at right angles to it in Figure 10. The comparison of the El Centro and Los Angeles results brings out some relations and similarities even though the localities are widely separated and geologically different. The longitudinal motion on the order of 3-second period in both cases rises to an early maximum and thereafter dies out very much as in the case of a more distant record. The short-period—3 to 5 seconds—transverse waves are not clearly in evidence, possibly because such waves appear to be related to direction of fault slip, and this is not very different from direction epicenter to station.

The existence of long-period surface waves in the epicentral area may be indicated by the presence of waves of 10-second period or more. This may indicate that such surface waves originate nearer the epicenter than Reid's (1931) view permits. The principal surface wave in this region is probably of the Love type—that is, transverse with no vertical component. In both records this wave can be readily identified and appears with almost the maximum displacement though the direction is normal to the fault. It is assumed that the vertical motion recorded at the same time is associated with other wave types.

Accordingly, while the evidence is favorable to the formation near the fault of the ordinary types of seismic waves, the results should be tested when additional records permit, and further attempts should be made to explain some of the motions which are not yet clearly understood.

## POSSIBILITY OF RECORDING PERMANENT DISPLACEMENT

Field investigations show the relative movement of the two sides of the fault slip at the surface, but we do not know whether one or both sides moved. We have reason to believe that if the records were complete from the start and stations were suitably placed with regard to the fault, this question could be settled by integration, provided instrumental performance were perfect. This would be a valuable supplement to precise geodetic measurement which gives the total permanent displacement but not the different steps if there are more than one.

It is probable that the great depth of apparently loosely consolidated sediments in the Imperial Valley accounts for the high accelerations and long duration of the record as compared to others. The high accelerations are restricted to short-period waves and are in all probability confined to layers above the basement rock, while the longer-period waves undoubtedly represent fundamental motion of the rock.

## DOMINANT PERIODS

The question of dominant periods has been investigated in Germany, Japan, and California by means of artificially induced vibrations and by study of earthquakes. Near Göttingen, Germany, a period of 0.3 second was dominant, and in Tokyo, Japan, periods of 0.3 and 0.4 second prevailed in differing parts of the city in the 1923 earthquake. A similar period, 0.33 second, outstanding in the accelerograms of the Imperial Valley earthquake, indicates that these are short-period high-acceleration waves.

However, no tests by artificially induced vibrations have been made in the Imperial Valley. In the Los Angeles region the use of a large shaking machine has revealed a number of dominant periods of which the following are examples: 0.085, 0.094, 0.120, 0.177, 0.188, 0.20**, 0.22**, 0.24**, 0.28**, 0.30**, 0.58, 1.1*, 1.4*, 1.7*, 2.3**, and 3.0 seconds. One asterisk indicates a well-defined period; two, an outstanding or very persistent period. This indicates complexity in the geological structure. Outstanding periods recorded on a near-by accelerograph during the Long Beach earthquake were 0.13**, 0.23**, 0.25, 0.35, 0.70**, 1.2 approximately, and 2.5 seconds approximately. There was obviously a strong tendency for many of the natural ground periods to become outstanding during the shock, but there were some exceptions due in part, perhaps, to the fact that the observation points were not identical.

## CONCLUSIONS

(1) The accelerations and computed velocities and displacements (oscillatory) of the earth particle at El Centro from an earthquake 7 miles away have been determined with satisfactory accuracy.

(2) The known types of earthquake waves are probably developed at or near the focus, which is a point or small area. The direction of the fault slip seems to have little effect in distorting them.

(3) Short-period waves of high acceleration are probably chiefly in the upper sediments, but the larger oscillatory displacements are in all probability also in the rock below.

(4) If the records were complete from the start, and instrumental performance good, large permanent displacements at the recording stations could be obtained from the accelerograms.

(5) The large number of dominant ground periods found in the Los Angeles basin is probably due to the complex geological structure.

(6) A practicable method is now available for dealing with new acceleration records as they are obtained.

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# PERMIAN LIME-SECRETING ALGAE FROM THE GUADALUPE MOUNTAINS, NEW MEXICO

BY J. HARLAN JOHNSON

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## ABSTRACT

During the course of field work in the Guadalupe Mountains of New Mexico and in certain parts of the Apache Mountains of West Texas, a variety of Permian calcareous algae were found. Thin-section examination of selected specimens from critical localities in these areas reveals many representative genera and species. One new genus and species are here presented together with descriptions of the most abundant and best preserved algal forms in the collection examined. Probable algal growths of pseudo-pisolitic character, which occur intimately associated with other algae, are also discussed.

The structures and relationships of the algae themselves, together with the presence of numerous associated organisms in the fossiliferous beds, facilitate reconstruction of the ecological conditions which apparently existed on and about the reefs during their construction.

## INTRODUCTION AND ACKNOWLEDGMENTS

The presence of algal material in the Permian limestones of the Guadalupe Mountains of New Mexico has been known for many years (Ruedemann, 1929, p. 1070). Sidney Powers called them to the writer's attention in 1931. They were briefly discussed by Pia in 1937 and again by Johnson in 1938.

The present study is based on material obtained from several sources over a period of years: (1) collected by parties of students from the Colorado School of Mines on their annual inspection trips between 1934 and 1940; (2) a collection sent in by J. E. Adams in 1938; (3) specimens collected by the author during the summer of 1939 when he was assisted by R. K. DeFord, J. E. Adams, R. King, K. Ferguson, E. H. Stevens, and C. A. Mantz. Altogether over 800 specimens and 150 thin sections were studied. A few specimens from the Apache Mountains of West Texas were included.

The type and figured specimens are in the writer's collection at the Colorado School of Mines.

The manuscript was practically completed, and the species named, in August 1940 when the author received a copy of a paper by Pia (1940) describing a number of the forms. This necessitated some revision in order to include Pia's names and some of his descriptions and comments. Unfortunately Pia's paper was short, contained no illustrations, and his descriptions were very brief. This made a comparison between his described species and the material treated here difficult. The author has endeavored to do full justice to Pia's work and his claims to priority.

The completion of this study was made possible by a grant from the Penrose Bequest of the Geological Society of America. The writer wishes to thank Mrs. M. E. Dorr for laboratory assistance, R. K. DeFord for field assistance and criticism of the manuscript, and E. H. Stevens for his photographic treatment of the specimens illustrated.

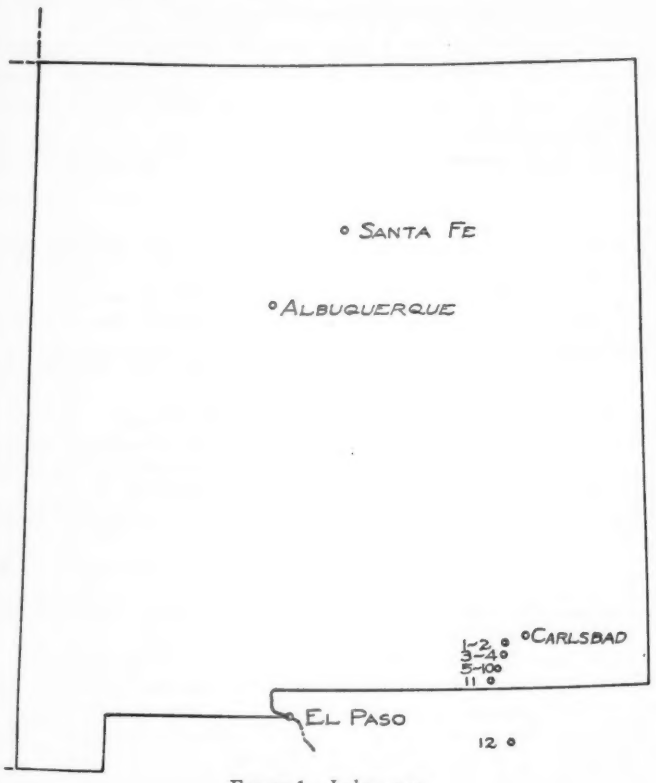


FIGURE 1.—*Index map*

## GEOGRAPHICAL DISTRIBUTION

The specimens were obtained from a number of localities in the Guadalupe Mountains and from one in the Apache Mountains of Texas.

The Guadalupe Mountains form a nearly north-south-trending range starting approximately west of the town of Carlsbad and extending

southward about 40 miles across southern New Mexico and over into West Texas. They consist of Permian sediments. Numerous ravines and canyons cut back into them and afford abundant outcrops for study. The relief of this range becomes progressively greater from north to south, culminating in the spectacular Guadalupe Point and Signal Peak at the southern end. The steep eastern scarp of the range represents essentially the dissected face of a Permian reef.

The Apache Mountains are a small range about 80 miles southeast of the southern end of the Guadalupe Mountains in Culberson County, Texas. The rocks are similar in stratigraphic range and lithology to those in the Guadalupes.

The individual localities are listed below and indicated on the map (Fig. 1).

LOCALITIES

1. Small shallow draw, approximately N. half sec. 21, T. 22 S., R. 26 E. Tansill formation, very near the top.
2. Draw 4 miles WSW. of Carlsbad, approximately NE. ¼ sec. 16, T. 22 S., R. 26 E. This is near the southeastern end of an escarpment about 5 miles long that faces northeast and trends southeast. The draw is the southernmost of those draining eastward. The escarpment forms the southwest side of the main "Hackberry Draw."
3. Dark Canyon, about 1 mile from mouth, north side of canyon.
4. Dark Canyon, about 1½ miles up canyon. South side of canyon on north side of divide between Dark and Moseley canyons. Yates formation between first and second Yates sands. Approximately W. ½ sec. 26, T. 23 S., R. 25 E.
5. Whites' City, slope to north, NE. ¼ sec. 34, T. 24 S., R. 25 E. Top Capitan (equivalent to top of Tansill).
6. Whites' City, along old road to Carlsbad Cavern in NW. ¼ sec. 34, T. 24 S., R. 25 E. Topmost Capitan.
7. Walnut Canyon, N. center SE. ¼ sec. 27, T. 24 S., R. 25 E. Capitan formation.
8. Walnut Canyon, E. side sec. 28, T. 24 S., R. 25 E. Tansill?
9. Walnut Canyon, NW. ¼ sec. 28, T. 24 S., R. 25 E. Tansill formation.
10. Carlsbad Cavern, about one-quarter mile SW. entrance, sec. 36, T. 24 S., R. 24 E. Tansill near top.
11. Rattlesnake Canyon, sec. 4, T. 25 S., R. 24 E. Capitan (Tansill age).

(Localities 1 to 11 in Eddy County, New Mexico.)

12. Apache Mountains, near Seven Heart Gap, Culberson Co., Texas.

## CLASSIFICATION

The forms described were classified on the basis of the following features: (a) microstructure when present; (b) size and shape of the colonies; and (c) growth habits. In description the highest forms are considered first, then progressively lower ones, ending with problematical types as indicated in Table 1.

The *Solenopora*, Dasycladaceae, Codiacea, and Porostromata show recognizable microstructure as well as definite shape of colony. The Spongiostroma are classified only on habit of growth and shape of colony. They may possibly represent form genera and form species rather than true biologic genera and species.

TABLE 1.—*Key to algal groups described*

| Class | Family-genus | Characteristic structures |
|---|---|---|
| Rhodophyta (Red algae) | *Solenopora* | Rows of cells so closely spaced that they assume polygonal cross sections. Cross partitions are present though frequently very thin. |
| Chlorophyta (Green algae) | Dasycladaceae<br>*Anthracoporella*<br>*Diplopora*<br>*Macroporella*<br>*Mizzia*<br>*Vermiporella* | A central stalk, showing as a tube or bulb, surrounded by tufts of leaves or leaf bases preserved as knobs or brushlike protuberances. |
| | Codiaceae | Small tubes loosely arranged so as to form segmented stems. Tubes round in cross section and branching. |
| Cyanophyta (possibly Chlorophyta) | Porostromata<br>*Girvanella* | Small tubes so loosely arranged that they do not compress each other. No cross partitions visible. |
| Cyanophyta (Blue-green algae) | Spongiostroma<br>*Cryptozoön*<br>*Ottonosia*<br>*Osagia* | Cellular structure seldom preserved. The lime is deposited as crusts on the outside of the colony or between the tissues, not in the cell wall. Classified on form of colony. |
| Problematical | Pisolitic masses<br>Crusts | Show little or no cellular structure. Have growth laminae suggestive of, but differing from, those of true pisolites (concretions) and lacking the crystalline structure of most chemical precipitates. |

Each of these form genera and species may represent several biologic species or even a number of genera and species which lived in constant close association. However, they seem to have been sufficiently persistent to be recognizable over wide areas through appreciable lengths of geologic time and to have formed colonies sufficiently different as to be easily separated by megascopic characteristics when entire colonies are available for study. Microscopic examination shows constant differences in the thickness and character of the growth laminae in different form genera, which seems to strengthen the validity of the classification.

## SYSTEMATIC DESCRIPTIONS

### Class RHODOPHYTA

### Family CORALLINA

### Subfamily MELOBESIEAE

### Genus *Solenopora* Dybouski

Specimens of the genus *Solenopora* were observed at many localities along the upper slopes and crest of the Guadalupian reef. Samples were collected in the vicinity of Whites' City and near the mouth of Dark Canyon. Unfortunately all the material obtained was poorly preserved.

#### *Solenopora centurionis* Pia

(Plate 1, figures 1, 3-6)

1937. *Solenopora* sp. PIA, p. 835, pl. 10, figs. 1-3.
1940. *Solenopora centurionis* PIA, p. 3.

Pia (1940, p. 3) gives essentially the following description: Thallus forms irregular, often lobate masses or nodules, 1 to 4 centimeters in diameter. Texture dense, usually concentrically banded with fine, dark lines between bands. In cross section the cells appear more rounded than angular. Originally the walls were probably thick, although this is clearly shown on only a few pieces, probably because of poor preservation. The diameter of the cell lumens is about 0.03 to 0.04 mm.; thickness of cell walls about 0.01 mm. or slightly more. No traverse partitions were found, possibly as a result of poor preservation.

The writer would add the following observations: Cells long and slender with uniform width over the short lengths measurable. No transverse partitions visible in longitudinal section. Transverse sections show the cells to be irregularly polygonal to hexagonal with straight sides. No septiform processes extend into the interior of the cell.

Darkened areas at the same level in the cell walls, which form gently arching, concentric lines, may possibly indicate growth stages. The distance between two "growth stage" lines in one relatively well preserved example was about 0.350 mm.

Cell measurements

| Width longitudinal section | Width transverse section | Wall thickness |
|---|---|---|
| .035–.070 | .035–.0525 | |
| .035 (approximate) | .035 (approximate) | |
| .043–.0525 | | .0175 |

CLASSIFICATION: The cellular structure is definitely that of a *Solenopora* in the restricted sense (Pia, 1939, p. 742).

REMARKS: This form, according to Pia (1940, p. 3), closely resembles *S. dionantina*. Chronologically it is closest to *Parachaetes triasinus* (Vinassa) but differs from it in the absence of transverse partitions, a larger cell size, and in forming larger colonial masses.

Only a few specimens were found in the collections studied. All were poorly preserved and perforated by animal borings.

In general aspect this species of *Solenopora* is fine textured. In most of the examples the outlines of the cells are somewhat obscured as if covered by a film

of cloudy, structureless material. *Girvanella*-like tubes of medium size were entwined along the "growth layers" of one specimen.

OCCURRENCE: Locality 5 near Whites' City and localities 2 and 3.

*Solenopora* sp.

(Plate 1, figure 2)

Fragments of another *Solenopora* having tubes of larger diameter than *S. centurionis* were found. The tubes measure between 0.065 and 0.095 mm. One slide shows definite cell division with dichotomous branching of one wall of the original cell. All available material was too fragmentary to justify naming the species although it seems to belong to the same one as that mentioned by Pia (1940, p. 3-4) as *Solenopora* spec. ind. His statement "tubes from 0.06 to 0.10 mm. in diameter, transverse partitions appear to be lacking" fits these fragments.

OCCURRENCE: Localities 2, 3, 4, and 5 along the east face of the northern end of the Guadalupe Mountains.

## Class CHLOROPHYTA

## Subclass CHLOROPHYCEAE

## Order SIPHONOCLADALES

### Family DASYCLADACEAE

GENERAL: The Dasycladaceae are nonpartitioned, multinucleate algae. The thallus is composed of a stem cell and a stem with branches arranged in whorls or verticillatae. The Dasycladaceae are strongly calcified, and the calcification includes either (1) the sporangia, (2) the cellulose membrane of the entire plant, which in this case is surrounded by a solid sheath of calcium carbonate reproducing all the details of the structure, or (3) both the sporangia and the membrane. (Morellet and Morellet, 1913.)

Algae of this type were abundantly represented in many of the specimens. They were the most interesting group studied and offer the greatest possibilities of any for recognition and use in petroleum geology.

Remains of the calcareous parts of Dasycladaceae were found in about a third of the thin sections from the Guadalupe Mountains near Carlsbad, New Mexico, and in the few obtained from the Apache Mountains, Texas. The state of preservation varied from those so recrystallized that only the outline of the organism remained to others in which it was possible to distinguish even fine details of internal structure. The material in most cases showed signs of wear and abrasion by wave action or similar agencies before burial in the sediments. In many cases the fragments received a finely banded concentric coating around the exterior which occasionally extended to the interior cavity as well. These banded coatings are often composed in part of tubular filamentous algae allied to *Girvanella;* in others the coating showed no organic structure and may be "dolomite deposited about them before they were buried in the calcareous sediments" as suggested by Girty (1908). The best material was obtained from Walnut Canyon (Fig. 1) and a locality in the Apache Mountains. Additional good material came from locality 2, south of the Cueva re-entrant near Carlsbad.

ASSOCIATED ORGANISMS: Of the associated organisms the filamentous algae are the most important both numerically and because of their close association with the Dasycladaceae. In addition to coatings they form numerous nodules and pellets and also occur in isolated fragments and groups of segments. They are of

varying dimensions and habits and will be described in detail in the discussion of *Girvanella* and the Cyanophyceae. The slender, nonpartitioned tubes, however, are those most often found in intimate association with the family under discussion. By their enlacings they form a constituent part of the coating so often observed around the Dasycladaceae.

Foraminifera of various genera, including a few fusulinids, are found in the sections with the Dasycladaceae. As a rule the fusulinids are not numerous in the slides containing abundant Dasycladaceae, and those containing many fusulinids show only rare and fragmentary Dasycladaceae.

Some Bryozoa fragments, in most cases apparently belonging to the species *Acanthocladia guadalupensis* Girty, were observed. Slender acicular rods interpreted as sponge spicules are numerous. Fragments of echinoderm plates are also relatively abundant. Some ostracodes are present but not numerous. There is also a wealth of indeterminate organic debris in most of the sections.

The Dasycladaceae in the slides from the Guadalupe Mountain region usually seem to be smaller than those forms described from Europe (Pia, 1920) although they closely resemble them in other respects. Their uniformly small size is a constant characteristic. It can hardly be interpreted as resulting from random, not equatorial or central, slicing of the plants in the section. The rather constant order of the size ratio in the specimens, which agree most closely with Pia's descriptions in other characteristics, seems to eliminate this possibility.

Table 2 shows selected dimensions of Pia's (1920) Carboniferous forms compared to similar dimensions for the types found in the American sections. All dimensions are in millimeters and were averaged for the different forms.

TABLE 2.—*Comparison of D, d for Pia's Carboniferous forms and Permian forms (mm.)**

| | D | d |
|---|---|---|
| Carboniferous of Europe (Pia) | | |
| *Anthracoporella* | 3.06 | 1.9 |
| *Mizzia velebitana* Schubert | 2.16 | 1.3 |
| *Mizzia yabei* Karpinsky | .75 | .38 |
| Permian-Guadalupe Mountain region | | |
| *Mizzia velebitana* Schubert "Cervical form" | .510 | .277 |
| *Mizzia yabei* Karpinsky | .877 | .449 |
| *Diplopora* (?) sp. | .696 | .333 |

The symbols used for the measurements, both in Table 2 and in the tables of representative measurements for the forms described later, are those of Pia with the addition of a few from Hφeg (1932). They are listed in tabular form below, an asterisk denoting those taken from Hφeg.

D—outer diameter of calcareous body
d—inner diameter
p—diameter of pores
g—number of verticillatae in a single member
h—distance between centers of verticillatae
st—diameter of stem cell
—inclination of the pores (angle made by pores with principal axis of the thallus)
*i—distance between center of pores on the surface
*s—thickness of calcareous wall

Too great reliance should not be placed on the measurements as indicating the true dimensions of the living plant, *i.e.*, the calcareous parts of the living plant, since many of the specimens were worn and abraded before burial. Measurements of specimens surrounded by banded coatings are even less representative of actual dimensions than the others since it is difficult to determine accurately the original outlines of the calcified parts of the plant, and here again considerable abrasion probably has taken place.

TERMINOLOGY: The terminology of the writers consulted varies somewhat. Høeg appears to use the term "pore," and Pia, verticillatae, to designate any opening in the outer calcareous wall. Morellet and Morellet distinguish between primary and secondary branches and sporangia. These distinctions probably resulted from the better state of preservation of their Tertiary material. The author uses the term "pore" in the same general sense as Høeg since the preservation of his material makes it difficult to distinguish between the different types of structure in most specimens.

CLASSIFICATION: The Dasycladaceae in the Permian sections appear to fall into two general groups:

1. Those specimens in which the thallus is composed of a series of spherical to lenticular members which were originally connected by a continuation of the stem cell. These may be referred to the genus *Mizzia*.
2. Those specimens in which the thallus was probably columnar to club-shaped and consisted of a single member. These are referred to the genera *Macroporella* and *Diplopora*, as are most of the other indeterminate specimens in the slides. Such forms appear to be the most common among all Dasycladaceae, fossil or modern.

## Genus *Mizzia* Schubert, 1907

Dasycladaceae with thallus composed of several spherical or elongated members growing on a common stem, suggesting a string of beads (Fig. 2). The members usually break apart at the narrow, connecting necks. Pores unbranched, increasing in size toward the exterior and arranged in concentric rows around the stem. The spores probably were produced in the central stalk.

*Mizzia* may be distinguished from the other genera of Dasycladaceae by its beaded structure, by the spherical form of the undivided members, and by the exceedingly simple structure of the branches as shown by the pores.

According to our present knowledge the genus *Mizzia* is limited to the Permian and reached its greatest development in the late middle and upper Permian. Geographically its distribution is practically worldwide. To date it has been found in southern Europe, Japan, the East Indies, west Texas, and New Mexico.

### *Mizzia velebitana* Schubert

(Plate 2, figures 1-4; Plate 3, figures 1-4)

1907. *Mizzia* SCHUBERT, p. 212.
1908. *Mizzia velebitana* SCHUBERT, p. 382, pl. 16, figs. 8-12.
1908. *Mizzia velebitana* KARPINSKY, p. 262, pl. 3, figs. 6-9.
1908. *Mizzia* cf. *velebitana* KARPINSKY, p. 266, pl. 3, figs. 1, 3, 4, 10-13.
1908. *Guadalupia* ? sp. GIRTY, p. 85, pl. 5, figs. 7-11.
1920. *Mizzia velebitana* PIA, p. 19, pl. 1, figs. 12-23.
1937. *Mizzia velebitana* PIA, p. 765, pl. 9, fig. 3.
1940. *Mizzia velebitana* PIA.
1942. *Mizzia velebitana* JOHNSON AND DORR.

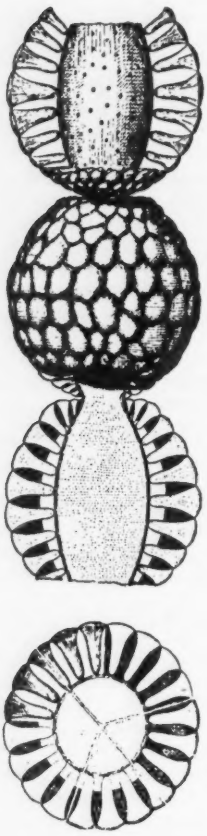

FIGURE 2.—*Pia's reconstruction of Mizzia velebitana*

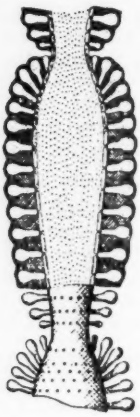

FIGURE 3.—*Pia's reconstruction of Mizzia yabei*

The thallus is composed of members with spherical or slightly ellipsoidal form with two rather wide polar openings opposite one another.

Some of the *Mizzia* studied fit the description and dimensions of this form very closely.

Measurements (Guadalupe Mountain specimens)

| | | |
|---|---|---|
| 1. | D—1.9 mm. | d—0.9 mm.—48% |
| 2. | D—2.0 mm. | d—st.—1.1 mm.—54% |
| | | H—2.2 mm.—107% |
| 3. | D—2.6 mm. | d—1.9 mm.—74% |
| | | st—1.3 mm.—51% |

REMARKS: The development of the calcified portion is unusually variable. It may extend inward as far as the stem cell, or more frequently the calcareous layer is thinner and has a very irregular interior border. In incompletely calcified specimens, instead of a thick wall, a thin calcareous layer may cover not only the stem cell but outline all the known important structures of the plant.

The verticillatae do not appear to be arranged in any constant pattern. They have a marked expansion toward the exterior and are somewhat polygonal in their distal parts as a result of mutual flattening.

The number of pores in the equatorial region is about 20 to 25.

In the central part of each sphere the pores are approximately perpendicular to the thallus. Their length and cross section decrease from this point in both directions. Their inclination to the long axis of the sphere increases rapidly toward the upper polar opening so that the last pores in that region are nearly parallel to the axis of the thallus, whereas from the equator to the lower polar opening the inclination is only slightly in the downward direction. The pores around the upper opening are smaller than on the other areas.

The central stem has the greatest diameter in the equatorial region of the spherical members, gradually tapering off in both directions to the polar openings.

No specimens have been found with openings on the side; hence the thallus of *Mizzia velebitana* is considered to have been unbranched.

OCCURRENCE: Localities 2, 3, 5, 7, 9, 10. Most abundant at 2, 9, and 10.

### *Mizzia minuta* Johnson and Dorr

(Plate 4, figure 4; Plate 5, figure 4)

1942. *Mizzia minuta* JOHNSON AND DORR, Jour. Paleont., vol. 16, p. 73, pl. 9, fig. 2; pl. 12, fig. 2.

Thallus unbranched and consisting of a series of spherical to ellipsoidal members developed along one axial cell. Pores unbranched with cross section of a truncated cone and usually open to the exterior; those which appear to be closed to the exterior have a more rounded cross section.

Number of pores in equatorial region from 13 to 16. Pores in the upper portion of the individual members have a position almost parallel to the main axis of the plant, those in the equatorial region are more nearly perpendicular to this axis, and basal pores are only slightly inclined toward the axis in a direction opposite from that of the upper pores. Pores usually open to the exterior but appear to be closed in one or two instances.

The walls are uniform in thickness throughout an individual member and bear a rather constant ratio to the dimensions of the plant, being from 0.140 to 0.263 mm. thick. Two examples showed exceptionally thin walls (about 0.085 mm.) but retained uniform thickness throughout the member. Most specimens showed a thick calcareous wall which probably included the area from the exterior to the stem cell. Two specimens show an outer thin calcareous wall terminating in an irregular border toward the interior, succeeded by open spaces in a second calcareous area and finally by the central cavity defined by a thin dark line.

REMARKS: The specimens of this form from the Guadalupe Mountains agree very closely with the description given for *M. velebitana* with the exception of their smaller size and smaller number of pores in the equatorial region.

In most of the specimens studied, the space within the outer calcareous wall is not divided into a central stem and other parts, the whole central cavity being composed of coarsely crystalline calcite or, occasionally, of fine-grained calcite. This

might indicate either (1) that the stem of the living plant occupied this entire area, or (2) that the walls of this cell received no calcareous coating during the life of the plant and were hence not preserved after the death of the organism, or (3)

TABLE 3.—*Representative measurements for Mizzia minuta Johnson and Dorr (mm.)*

| D* | d** | st | s | p | i | No. of pores equatorial region |
|---|---|---|---|---|---|---|
| .875 | .263 | | .263 | | | |
| 1.225 | | | .263 | | | |
| | | | .350 | | | |
| | | | .437 | | | |
| .752 | .402 | | .175 | .052 | | 14 |
| .700 | | | .157 | .087 | | |
| | | | .140 | | | |
| .700 | .665 | | .087 | .087 | .157 | |
| | | | | .122 | .140 | |
| | | | | | | 15 |
| .962 | | .350 | .122 | | | |
| 1.312 | | .525 | .157 | | | |
| .962 | .350 | | | | | |
| .700 | .490 | | | | | |
| .875 | .437 | .227 | .175 | | | 13 |
| .900 | .350 | .175 | .210 | | | |
| .700 | | | .175 | .035 | | |
| .612 | .385 | | .175 | .052 | | |
| | .315 | | .140 | | | |
| | | | .122 | | | |
| .875 | .408 | | .175 | .700 | | |
| .700 | .525 | | .152 | .875 | | |
| | | | .140 | .875 | | |
| | | | .140 | .105 | | |
| | | | .122 | | | |
| .875 | .437 | | .175 | | | |
| | | | | | | 16 |

* Average .795 mm.
** Average .418 mm.

that the stem, being attached to the outer calcareous part by slender branches or canals which would break off readily, might drop out of the broken fragment during agitation by waves and before burial in sediments.

*Mizzia minuta* is by far the most abundant species of this genus in the area studied.

OCCURRENCE: Localities 2, 8, 9, 10, 12. Abundant at 2, 9, and 10.

*Mizzia yabei* (Karpinsky)

(Plate 3, figures 4, 6; Plate 7, figure 2)

1908. *Stolleyella* (?) *yabei* Karpinsky, p. 268, 269, figures.
?1908. *Stolleyella velebitana* Karpinsky, pl. 3, fig. 8.
1920. *Mizzia yabei* Pia, p. 23, pl. 1, figs. 4-6.
1937. *Mizzia yabei* Pia.
1942. *Mizzia yabei* Johnson and Dorr.

Elongate, elliptical form, about twice as long as wide. Pores and members unbranched. As with *M. velebitana,* several members were probably attached to a communal stem, with the limits between the individual members indicated by a marked constriction. The thallus breaks up into the individual members and fragments thereof upon the death of the plant. Broken fragments show the constricted area most frequently as a "neck-like" structure often open at the end. No examples were observed wherein the two members are shown connected.

Table 4.—*Measurements of specimens from the Guadalupe Mountains (in mm.)*

| D* | d** | s | p |
|---|---|---|---|
| 1.120 | .525 | .265 | .052 |
| | | .297 | .087 |
| | | | .087 |
| | | | .122 |
| .875 | .315 | .263 | |
| .910 | .577 | .175 | |
| .875 | .472 | .210 | |
| .787 | .437 | .175 | |
| .700 | .350 | .175 | .14 (approximate) |
| | | .263 | |
| .875 | .472 | .210 | |

* Average .877 mm.
** Average .449 mm.

Remarks: *M. yabei* differs from *M. velebitana* in its smaller size and in having members with an elongate oval form. Differs from *M. minuta* in the ovate form of the members.

Pia (1920) gives an instructive reconstruction showing his idea of the structure. He admits that it is very problematical as many important details of the structure were unknown (Fig. 3).

Transverse sections of *M. minuta* and *M. yabei,* particularly those cut in the equatorial region, must be so similar as to render it extremely difficult to distinguish between them. The same, except for size, is true of *M. velebitana.* A few of the very small transverse sections observed but not classified may represent sections through *M. yabei* at a point within or near the constricted area.

This form is less abundant and, generally, less well preserved than the other species of *Mizzia* in the Guadalupe Mountain region. Broken members or those more or less tangential to longitudinal sections of the upper and lower ends of the individual members are most frequent, and entire members are rare.

The symmetrical outline of the individual members constitutes a divergence from Pia's European forms. Pia tentatively states that, in general, his forms show the maximum diameter in the upper third of the member. Only one of the New Mexico examples shows this characteristic. However, the statement made by Pia is not definitely expressed, and such minor variations should not prevent reference of these specimens to this species.

The difficulty of distinguishing between transverse sections of *M. minuta* and *M. yabei* made it impossible to ascertain with assurance the number of pores in the equatorial region of the latter.

The pores have a marked expansion toward the exterior and have the outline of a truncated cone. Pia shows the pores terminating within the calcareous wall in a very rounded end in contrast with the open form found in the Guadalupian specimens. Too great importance should not be attached to this divergence inasmuch as, although Pia's reconstruction of *M. velebitana* shows widely open pores, one of his plate figures for the same species depicts a pore termination very similar to that which he illustrates for *M. yabei*.

The position of the pores is generally perpendicular to the principal axis of the thallus at the equatorial region, with a steeper inclination at the top and a very slight one in the opposite direction at the basal portion.

The pores are almost always open to the exterior. Where occasionally they seem to be closed, this appearance may be due to the tangential nature of the section.

The calcareous outer wall is usually relatively thick and uniform throughout the member. No examples of "incomplete calcification" were observed.

The stem cell has an apparently central position, often seeming to fill the entire central cavity, although this point cannot be definitely confirmed with specimens as recrystallized as those studied.

No reproductive organs or traces thereof were observed.

The constriction at the "neck" in at least one specimen was rather abrupt, but this appearance may very possibly have been due to breakage. Generally the constriction is more gradual.

OCCURRENCE: Localities 5 and 9.

## Genus *Macroporella* Pia 1912

### *Macroporella* cf. *verticillata* Pia

1908. *Guadalupia* ? GIRTY, p. 86, pl. 27, figs. 11-12.
1940. *Macroporella verticillata* PIA, *p.* 2.

Thallus cylindrical to barrel-shaped. Pores expand toward the exterior, sometimes only moderately, sometimes strongly. They may change from one extreme to the other in the same plant.

Dimensions

| | *a* | *b* |
|---|---|---|
| Exterior diameter of test..... | 0.88 mm. | 1.15 mm. |
| Interior diameter of test..... | 0.50 mm. | 0.48 mm. |
| Interval between whorls..... | 0.13 mm. | 0.16 mm. |

REMARKS: Only one other *Macroporella* with whorls has been described in detail to date—the upper Jurassic *M. pygmaea* Gümbel (Pia, 1924, p. 84). This species is smaller than *M. verticillata* with thicker walls, and the terminations of the verticillatae form a complete crustal layer. *M. tetrapora* Pia (1937, pl. 10, fig. 4) from the Permian of Sumatra differs from *M. verticillata* in its greater size and in other characteristics which it is hoped will be described more fully by Pia in the future. The latter's brief and inadequate description of the species and absence of illustra-

tions make it difficult to identify in the present work. No forms were found, however, which seem to fit closely the available description.

OCCURRENCE: Pia gave no locality beyond upper Permian of Guadalupe and Delaware mountains, New Mexico, and Texas.

*Macroporella* sp. ?

(Plate 4, figures 1, 2; Plate 3, figure 5; Plate 7, figure 4)

Fragments of a *Macroporella* were found which do not seem to belong to any described species. The material, however, was so poor and fragmentary that it did not seem advisable to name the species.

OCCURRENCE: Locality 8, Walnut Canyon, and Locality 6 near Whites' City.

Genus *Anthracoporella* Pia 1920

*Anthracoporella* sp. ?

(Plate 5, figure 5)

A single transverse section of a member of this genus was observed. It agrees closely with the description and reconstruction by Pia (1920, p. 15) of the genus as far as could be determined from one section. It had a diameter of 0.560 to 0.70 mm.

To date only one species, *A. spectabilis* Pia, has been described from the Permian. This occurred in the lower Permian of the southern Alps. The New Mexico specimen is quite similar but much smaller.

It is not considered advisable to name the form on the basis of such inadequate material.

OCCURRENCE: Locality 8, Walnut Canyon.

Genus *Diplopora* Schafhäutl 1863

*Diplopora* sp. ?

(Plate 3, figure 5; Plate 4, figure 3; Plate 7, figure 4)

Thallus long and slender, tapering markedly to acicular termination. Pores slender, probably branched, usually opening to exterior. Thallus usually shows a slight constriction and swelling toward the distal portion. The general outline is gently undulatory or irregular. On the lower third of the thallus the exterior openings of the pores have a characteristic "hooked" appearance. Longitudinal sections, including the central cavity, show a characteristic, almost bilobate development of the cavity at one end. Longitudinal sections usually show three well-marked pore openings, suggesting that the thallus bore six to eight branches in each whorl.

Pores are slender at base, expanding abruptly about midway along their length. The expanded portion is subrectangular or subcircular in outline. This expansion is interpreted as indicating branching into two or three branchlets with bulbous or subrectangular bases. The pores are spaced at regular intervals along the thallus. They appear to have been developed approximately perpendicular to the main stem which formed the central cavity in the proximal and distal portions of the thallus, but to be inclined in the equatorial region. This causes the "hooked" appearance.

The outer calcareous wall is relatively thick and of uniform dimensions throughout the thallus. No specimens showing incomplete calcification were observed.

REMARKS: *Diplopora* and *Macroporella* are very closely related structurally. Pia (1927) puts them together in the Tribus Diploporeae but is rather vague on the features which separate them. No species have hitherto been described from rocks older than Triassic. Specimens considered to be *Diplopora* were abundant at one horizon in Walnut Canyon and near Whites' City. Unfortunately the material

was poorly preserved. Several species seem to be present, but until better specimens can be obtained it does not seem wise to give specific names.

## Class CYANOPHYTA (possibly CHLOROPHYTA)

### Family Porostroma

### Genus *Girvanella* Nicholson and Etheridge 1880

(Plate 5, figures 1-3)

General: Filamentous algae, in the form of simple tubules, were present in more than half the thin sections. They did not exhibit a very marked development

Table 5.—*Representative measurements for Diplopora (?) sp. (mm.)*

| D* | d** | st | s | p | i | h | No. pores equatorial region |
|---|---|---|---|---|---|---|---|
| .612 | .175 | | .175 | .070 | .263 | | |
| .700 | .263 | | | | .227 | | |
| .752 | .385 | | .175 | .052 | .227 | | |
| | | | | .070 | | | |
| | | | | .052– | .263 | | 6 |
| | | | | .070 | .280 | .175 | |
| .875 | .350 | | | | | | |
| .787 | | | | | | | |
| .875 | | | | | | | |
| .105 | | | | | | | |
| | | .087 | | | | | |
| .910 | .437 | | .175– | .175 | | | |
| | | | .265 | | | | |
| | | | | .070– | | | 6 |
| | | | | .087 | | | |
| .612 | .265 | | | | | | |
| .525 | | | | .052– | .210– | .210– | |
| .612 | | | | .070 | .575 | .175 | |
| .700 | | | | | | | |
| 1.050 | .525 | | .263 | | | | |
| .962 | .408 | | | | | | |
| .875 | .385 | | .297 | | | | |
| .350 | .263 | | | | | | |
| .612 | .263 | | .157 | | | | |
| .700 | .350 | | .175 | | | | |
| .612 | .263 | | .175 | | | | |
| | | | .263 | | | | |

* Average .696 mm.
** Average .333 mm.

from any one locality nor did they form colonies of any great size. The tubes formed small irregular masses and coatings. Only rarely were well-defined nodules developed.

DESCRIPTION: It has been customary to classify and name the *Girvanella* on the basis of tube dimensions. In the Guadalupian material the width of the tubes varied within wide limits, and in numerous instances tubes of different dimensions were closely entwined with one another. However, they could be roughly divided into three classes: fine tubes averaging about 0.003 mm. in width, medium ranging from 0.0175 to 0.03 mm., and coarse tubes having a width of about 0.0525 mm. Transitional sizes were present, especially between the fine and medium groups. Medium-sized forms were the most abundant. The coarse ones were the least numerous. Although certain other structural features appeared to be characteristic of one or the other size groups, too many exceptions were found to give a strong basis for classification.

Transverse partitions were observed in only a small percentage of the tubes. They were most common in the medium-sized ones but were observed in others. In the medium tubes the partitions were straight, showing no arching as they crossed from one side to the other. These medium tubes with partitions are reminiscent of *Symploca jurassica* (Frémy and Dangeard, 1935).

The majority of the tubules appear to be of uniform diameter. Swellings and constrictions are rare. A few tubes with transverse partitions show distinct swellings, however. These occur just below the partitions. This feature is extremely rare and seen only in a few of the very fine tubes.

Definite branching is found in a few rare instances among very fine tubes.

The tubules of all diameters are outlined by fine, dark lines. There are no suggestions of sheath. Although a transverse section of the individual tube usually shows a fine, dark line defining the simple circular outline, numerous examples also show a darker, rounded central area within. This is most common among medium-sized tubes.

Clearly defined, thick walls were observed only twice where the thickness of wall could be measured. In one, an isolated segment of a large tube, the wall measured 0.175 mm., and the tube including the walls had a width of 0.0525 mm. The second, a segment of a smaller tube, showed a wall thickness of 0.009 mm., and the tube including walls about 0.030 mm.

It is not advisable to name these three types. Too many *Girvanella* have been defined already on inadequate material.

### Other tubular algae

Two other types of tubular algae were observed which differ sufficiently from the *Girvanella* to be considered as belonging to different groups.

The first and numerically most important group consists of very fine tubes with diameters ranging from 0.001 to 0.013 mm. These occur either perforating the test of some larger organism or penetrating a dark mass of organic debris. This form shows frequent dichotomous branching at acute angles.

The second type includes a few examples of large tubes having a peculiar, flattened, ribbon-like appearance. Except for the flattening they resemble the coarse *Girvanella*.

Class CYANOPHYTA ?

Family SPONGIOSTROMA

Genus *Collenella* Johnson, n. gen.

GENOHOLOTYPE: *Collenella guadalupensis* Johnson, n. sp.

Calcareous algae which form compact, compound colonies consisting of irregular columnar growths. Laminae of variable thickness form convex growth layers approximately perpendicular to axes of the columnar colonies.

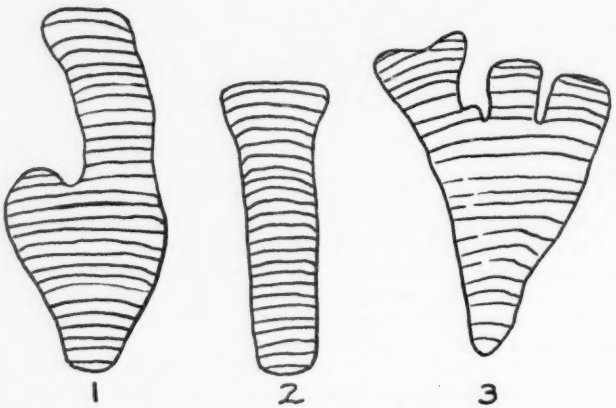


FIGURE 4.—*Growth habits of Collenella guadalupensis Johnson*

*Collenella guadalupensis* Johnson, n. sp.

(Plate 7, figure 3)

Colonies irregularly columnar, expanding slowly upward where not too tightly surrounded by other colonies. Have a plumose appearance in vertical section (Pl. 7, fig. 3). Growth laminae vary in thickness from 0.1 to 0.3 cm., averaging about 0.2 cm. Columnar colonies 3 to 14 cm. long, average width about 1.6 cm., maximum width observed 3.5 cm. Colonies may branch (Fig. 4) or stop and then be covered by a cluster of small ones. The longer colonies tend to pinch and swell more than the shorter ones.

Under the microscope (at x50) the laminae are composed of filaments forming an irregular mat. Some particles of foreign material are enclosed by the filaments, whereas around the margins of the laminae particles of extraneous matter may be present in considerable abundance. Hollow spaces within and between the laminae have been filled with clear crystalline calcite.

REMARKS: Named from the Guadalupe Mountains of New Mexico. The compound colonies form masses up to a foot or 18 inches in diameter.

CLASSIFICATION: The microstructure and growth habits suggest some of the higher types of the Cyanophyta or possibly a low form of the Chlorophyta.

OCCURRENCE: Locality 4 on south side of Dark Canyon, on the north side of divide between Dark and Moseley canyons. Stratigraphically between the first and second sands of the Yates formation.

## PROBLEMATICAL FORMS

### Pisolitic masses

(Plate 6, figures 1-3)

Pisolitic limestones are very common in the Guadalupe Mountains, especially in the lagoonal facies behind the reefs. The pisolites in these areas were early considered of algal origin (Ruedemann, 1929). This interpretation has been repeatedly questioned. Pia (1940) says they are inorganic. Many were studied in the present work in both hand specimens and thin sections.

As typically developed the "pisolites" are spherical or subspherical with flattened base and top. The average size is from 0.6 to 1.1 cm., and in many deposits they are surprisingly uniform. In very rare cases they grow much larger, and smaller ones also occur. They are formed of thin layers of material more or less concentrically arranged around a nucleus. In small "pisolites" and in the central portions of larger ones the layers are concentric, completely enveloping those beneath. As the object becomes larger, however, the layers tend to envelop the mass only partially, and the "pisolites" become flattened and rudely elliptical in cross section. The individual layers are seldom of uniform thickness. The irregularity becomes more pronounced in the outer layers where they thin out toward the margins. The nucleus may be a small gastropod, foraminifer, a segment of a Dasycladaceae, or a fragment of some other fossil; only rarely is the nucleus of inorganic material.

Microscopic examination shows the layers to be composed of very fine particles of calcium carbonate in most cases. Some show a definitely crystalline structure. This is considered as probably secondary since it is best developed in the outer layers. Even under high magnifications no cellular structure could be definitely observed although vague suggestions of a feltlike mat of filaments were occasionally found. A few *Girvanella* tubes were observed in some of the "pisolites." However, these were also observed in about the same abundance in most of the other objects studied, so their presence is considered more or less accidental.

In some localities small "pisolites" occur which show a structure of fine, radiating, needlelike crystals with concentric layers absent or poorly developed. Except for size these are like the small oölites present in many of the specimens. These are interpreted as of inorganic origin.

ORIGIN: The writer believes that the majority of the "pisolites" are of algal origin, representing calcareous material deposited around the outer (growing) layers of colonies of low types of blue-green algae and fine silt and organic debris caught in the outer growing layers of such colonies. The structure suggests growth layers which start as a coating about a small object and grow concentrically until the colony becomes large enough to cut off the light from the basal portion, which, as the mass becomes larger, is more and more likely to be partly buried in the fine debris of the lagoon bottom. It may be that from time to time they were rolled over by tides or storm waves or other causes, and growth continued on the upper surfaces. In appearance these "pisolites" are very similar to the modern algal balls and biscuits forming in modern freshwater deposits (Clarke, 1900; Roddy, 1915). Perhaps the environment of deposition was very similar to that described on the top and back of a modern Hawaiian reef by Pollock (1928, p. 12). Discussing nodular algal colonies he says:

"growing on the nearby reef flats between the reef rim and the shore are many different kinds of nodules, and these are exceedingly abundant on the part of the living reef off shore from the Wailupe Fossil Reef. These nodules lie in the shallow depressions over almost the entire distance from the shore to reef rim, which had

here a width of about 1,200 feet (365 meters). They are so abundant that in many depressions they are in contact with one another over the whole bottom or piled up several deep, even where the bottom has a considerable amount of fine mud which is stirred up at every high tide and settled temporarily at every low tide. The mud is so abundant on the shoreward third of the Wailupe Fossil Reef that some of the nodules of algae are completely buried in it at low tide, yet many of them are alive, being reddish or purple. In this locality, and a few others, at every high tide the water on the reef flat is exceedingly turbid."

The algae forming the Hawaiian colonies are of different types (Rhodophyta) than those presumed to have been responsible for the "pisolites" under discussion. However, it has long been recognized that the growth form of algal colonies is largely a matter of environment, different groups developing similar-shaped colonies under similar environmental conditions (Pia, 1928).

Another fact supporting the probable organic origin of these "pisolites" is the presence of crustlike growths of similar material and transitional forms between such crusts and the pisolites in these same lagoonal deposits (Pl. 6, fig. 1).

CLASSIFICATION: The absence of preserved microstructure and the analogy in appearance and structure to modern algal biscuits formed largely or entirely by members of the Cyanophyta suggest that they probably belong with that class.

OCCURRENCE: Abundant in the lagoonal facies of deposits behind the main reef at all localities studied in the Guadalupe Mountains. Most of the specimens came from Localities 1, 2, 5, 8, and 9.

### Crusts

(Plate 6, figure 1)

Commonly associated with the "pisolites" and rarely with the other forms small crusts occur. Usually thin, they may develop into extensive bands or ribbonlike masses.

DESCRIPTION: Thin, flat, or irregular crustlike masses. Size usually from 0.6 to 3 cm. but occasionally much larger. At one locality (No. 2) they formed masses 3 to 5 cm. thick and over a meter long. Some develop on or around a bit of foreign substance; others seem to have spread out on the bottom of the lagoon; still others grew over the reef surface.

Microscopically they consist of thin, irregularly developed laminae similar to those found in the "pisolites." The laminae are composed of very fine particles of calcium carbonate with occasional larger fragments of foreign material. Little or no definite organic structure can be observed, although some suggestions of fine filaments are present, and algal tubes occasionally penetrate the crusts.

REMARKS: These are most abundant and best developed in the lagoonal deposits associated with abundant "pisolites."

Thin coatings of similar material were observed on fossils at a number of localities.

CLASSIFICATION: They are interpreted as deposited by members of the Cyanophyta partly as a result of their normal physiological absorption of $CO_2$ from water. Some of their substance may also be derived in part from fine particles of material becoming entangled in their filaments and gelatinous coating.

OCCURRENCE: At all the localities studied, but especially Nos. 1, 2, 5, 6, 7, and 8.

## ALGAL DEPOSITS

Most of the rock in which these algal remains occur is a light-colored dolomitic limestone. Usually it contains an abundance of small organic

remains and fragments and a greater or less amount of very fine calcareous particles of undetermined origin. The percentage of these constituents varies from bed to bed and from place to place. Similarly the deposits have suffered varying amounts of alteration, mainly solution and recrystallization. At some localities certain beds have been recrystallized to a porous dolomite which shows little or no organic structure. At other places so little alteration has occurred that details of structure of the organic fragments can be clearly seen even at moderately high magnifications. Such alteration is irregular and very variable even through short distances in the same bed.

Some horizons are predominantly organic, Foraminifera and algae being the commonest constituents. *Fusulina* limestones occur, especially in the lagoonal facies. Other laminae are composed largely of fragments of Dasycladaceae, especially *Mizzia*, forming *Mizzia* limestone (Pl. 2, figs. 3, 4; Pl. 4, figs. 1, 4). The "pisolitic" layers are abundant and thick in the lagoonal deposits (Pl. 6, figs. 2, 3). In other instances small Foraminifera and algal fragments may form a fine-textured chalky rock.

The main reef rock itself was built by algae, bryozoans, sponges, foraminifers, brachiopods, small mollusks, hydroids, and other organisms, but was very porous and is usually highly recrystallized. The best preserved material comes from the lagoonal facies close to the back of the reef.

Some of the rock resembles a calcium carbonate "mud" of very fine calcareous particles and shows very few organic structures even though not recrystallized. Those present are fine and medium algal tubes and occasional calcareous plates of holothurians. It probably was originally an algal mud such as is now being formed in the coastal lagoons of the Bahamas (Black, 1933) possibly somewhat reworked by the Holothuroidea.

The rocks range from slightly dolomitic limestones to pure dolomites. In the lagoonal facies some clastic material is present. This usually occurs as disseminated particles near the back of the reef but becomes more abundant shoreward until definite lenses of detrital material interfinger with the calcareous deposits.

Algae were probably responsible for much if not all of the magnesium carbonate present in the dolomite as magnesium is an essential constituent in chlorophyll, that green pigment characteristic of plants. Modern lime-secreting algae contain large quantities of magnesium carbonate as well as calcium carbonate (Howe, 1933; Clarke and Wheeler, 1922, p. 48-55). The writer believes that the importance of these plants in relation to the origin of dolomite has not been appreciated by geologists.

## ECOLOGY

Ecologically all the material sudied came from either some portion of the reef facies of these deposits or from that portion of the lagoonal facies adjoining the reef.

Four ecological facies are recognized. These are shown diagrammatically in Figure 5. They are (a) the reef face or front, (b) the reef crest or ridge, (c) the reef platform or flat, (d) the reef side of the lagoon.

[A] The reef face is a steep, rough slope along the sea side of the reef. Its inclination ranges from nearly vertical to 25 or 30 degrees. It is of

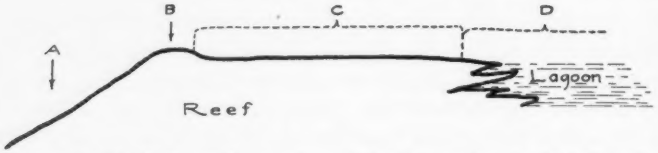


FIGURE 5.—*Idealized profile across reef showing various ecological forms*

massive, frequently porous, dolomitic limestone with bedding poor or absent. Organisms present are filamentous green and possibly blue-green algae with red algae becoming abundant near the crest, numerous Bryozoa, some corals (mainly simple horn types), some sponges, Hydrozoa, small gastropods, and a few small burrowing pelecypods. The algae were numerically the most important and not only were the greatest contributors to the rock material but seem to have acted as a binder to hold it together, much as the encrusting red algae do on the modern reefs of the tropical Pacific.

[B] The reef crest is usually developed as a low, narrow ridge at the top of the reef face, rising slightly higher than the reef platform behind. The rock resembles that of the reef front except that it is composed almost entirely of encrusting algal material (green and red) with colonies of *Solenopora* locally abundant on the inner slope. Fragments of Dasycladaceae occur. This was the portion of the reef exposed to the violent action of the waves and appears to have corresponded to the *Lithothamnium* ridge so frequently mentioned as forming the outer surface rim of modern reefs (Mayor, 1924; Pollock, 1928).

[C] The reef flat or platform represents the flat top of the reef behind the reef crest. By analogy with modern reefs it was probably partially exposed at low tides and was only covered by very shallow water at other times. The width differed greatly at different localities, ranging from a few hundred feet to around 1800 feet. The surface appears to have had some relief near the front where numerous pools and channels were

developed. The rocks are dolomitic limestones, poorly bedded, usually dense. The principal organisms observed were encrusting and filamentous algae, colonies of *Solenopora,* Bryozoa, abundant Foraminifera, some large gastropods, thick-shelled brachiopods and pelecypods, and holothurians. An abundance of organic debris, much calcareous mud, and some fine silt are present. Algal pisolites occur in some of the pools. At the rear the beds grade into or interfinger with typical lagoonal deposits.

The lagoonal facies extended from the reef to the shore, which was frequently miles away. Several ecological facies were differentiated in the lagoon itself. The algal deposits are largely confined to the reefward side where calcareous deposition predominated. The rocks are usually well stratified with thin to medium bedding. "Pisolites," "pisolites" and small crusts, and foraminiferal mixtures are common. Locally layers and lenses of fusulinid limestones occur, frequently with the *Fusulina* all having the same orientation. Occasional layers or patches of chalky limestone were observed and streaks and patches of calcareous muds are abundant. Shoreward these grade into or interfinger with clastic sediments. The organisms include an abundance of algae (green and probably blue-green, Dasycladaceae, fragments of *Solenopora*), abundant Foraminifera both large and small, a few ostracodes, small gastropods, worms (borings), and fish teeth. Naturally some materials originating in one facies may be found in others as a result of wave action; therefore the forms listed may not all have inhabited the areas in which they were found.

## CONCLUSIONS

Lime-secreting algae were abundant in the Guadalupe region during upper Permian time and were important in depositing dolomitic limestones. The reef limestones were formed largely by various types of algae, assisted to a lesser but locally varying extent by other organisms.

Much of the algal structure originally present has been lost by poor preservation, solution, and recrystallization of the deposits and by the activity of boring organisms. At least some of the dolomitic muds represent algal deposited material disintegrated by holothurians and other organisms. Most of the "pisolites" were probably deposited by low forms of algae although some are chemical precipitates.

Further studies will undoubtedly show numerous other examples of algal deposits in West Texas and adjoining regions and will probably lead to the recognition of many new forms.

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Colorado School of Mines, Golden, Colo.
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Project Grant 291/39.

## Explanation of Plates

### Plate 1.—*Solenopora*

Figure

1. *Solenopora centurionis* Pia (1937). A small colony, natural size. Guadalupe Mountains, New Mexico (p. 200).
2. *Solenopora* sp. (x1). Apache Mountains, Texas (p. 201).
3. *Solenopora centurionis* Pia (1937). Longitudinal section, x46 (p. 200).
4. *Solenopora centurionis* Pia (1937). Tangential section, x46 (p. 200).
5. *Solenopora cf. centurionis* Pia. Longitudinal section, x60 (p. 200).
6. *Solenopora cf. centurionis* Pia. Oblique section, x60 (p. 200).

FIGURE 1

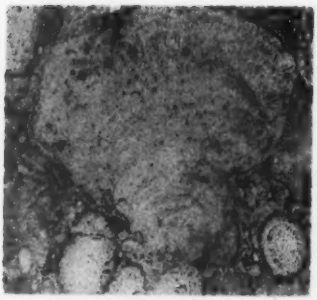

FIGURE 2

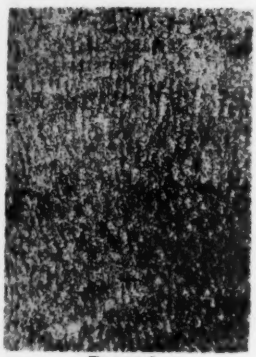

FIGURE 3

FIGURE 5

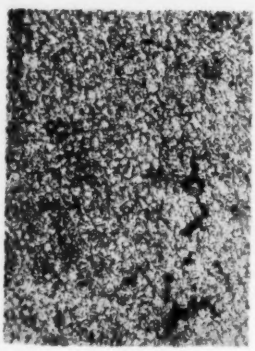

FIGURE 4

FIGURE 6

*SOLENOPORA*

FIGURE 1

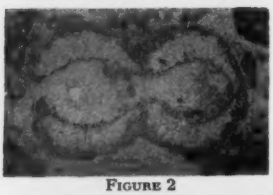

FIGURE 2

FIGURE 3

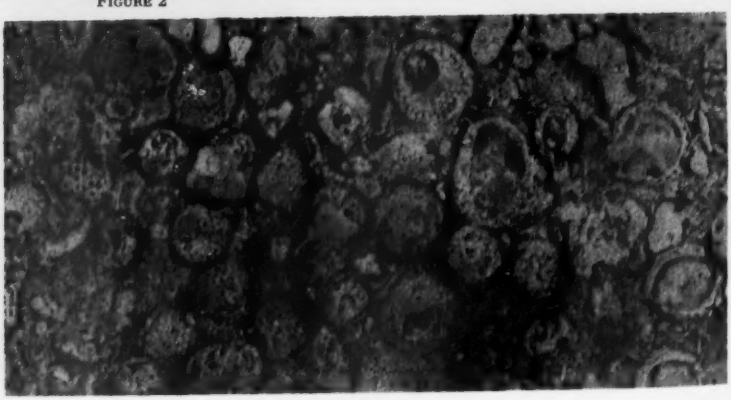

FIGURE 4

PLATE 2.—*Mizzia*

Figure

1. *Mizzia velebitana*. Showing characteristic weathering, polar openings, connected members, with elongated neck between, x4. Apache Mountains, Texas (p. 203).
2. *Mizzia velebitana*. Longitudinal section of two connected members, x25. Walnut Canyon, New Mexico (p. 203).
3. Typical "*Mizzia* limestone," formed mainly of *Mizzia velebitana*, x0.8. Apache Mountains, Texas (p. 203).
4. *Mizzia velebitana*. Enlarged detail of specimen shown in Figure 3, x8 (p. 203).

PLATE 3.—*Mizzia, Diplopora ? or Macroporella ?*

Figure

1. *Mizzia velebitana.* Tangential section. Characteristically abraded material, x25. Walnut Canyon, New Mexico (p. 203).
2. *Mizzia velebitana.* Longitudinal section showing character of wall and outlines of "pores." Also a tangential section of same species, x25 (p. 203).
3. *Mizzia velebitana.* Chain of small members, x7 (p. 203).
4. *Mizzia yabei* and *M. velebitana.* Sections at various angles. Some segments coated with dark layers of filamentous algae, x25 (p. 203).
5. *Diplopora* sp. ? or *Macroporella* ? Longitudinal section, x25 (p. 209).
6. *Mizzia yabei* (Karpinsky). Longitudinal and oblique sections, x25 (p. 207).

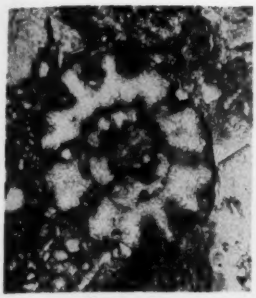

FIGURE 1

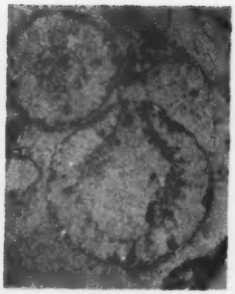

FIGURE 2

FIGURE 3

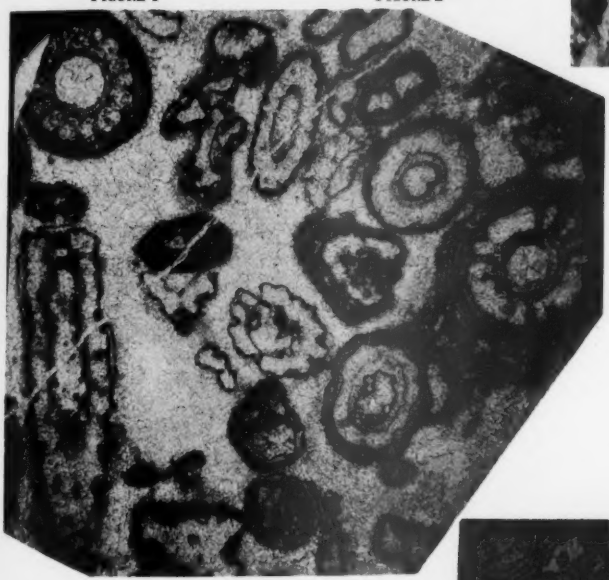

FIGURE 4

FIGURE 5

FIGURE 6

*MIZZIA, DIPLOPORA?* OR *MACROPORELLA?*

FIGURE 1

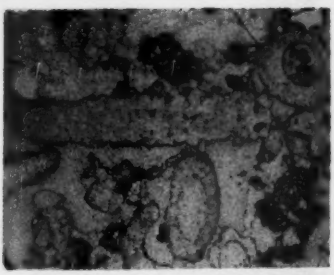

FIGURE 2

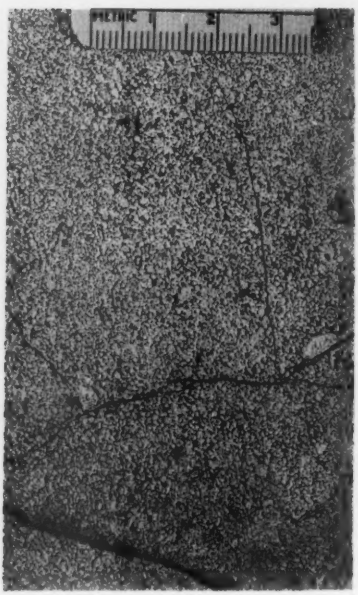


FIGURE 4

FIGURE 3

*MACROPORELLA, DIPLOPORA, MIZZIA*
Calcareous Algae, Guadalupe Mountains.

PLATE 4.—*Macroporella ? Diplopora, Mizzia*

Figure

1. Limestone composed largely of *Macroporella* sp., x0.8. Walnut Canyon, New Mexico (p. 209).
2. *Macroporella* ? sp. ? Longitudinal section showing general aspect and the highly recrystallized character of most of the specimens of *Diplopora* and *Macroporella* obtained in the Guadalupe Mountains, x10 (p. 209).
3. *Diplopora* ? sp. ? Longitudinal section, not central, showing characteristic arrangement and outline of pores, x10. Transverse sections of *Mizzia* also present (p. 209).
4. *Mizzia* limestone formed of *Mizzia minuta* Johnson and Dorr, x7. Apache Mountains, Texas (p. 205).

PLATE 5.—*Girvanella, Mizzia, Anthracoporella*

Figure

1. Tubular *(Girvanella)* algae entwined in finely concentric layers surrounding and filling much abraded transverse section of *Mizzia velebitana*, x50 (p. 210).
2. *Girvanella* filaments filling central cavity of a transverse section of a *Mizzia*, x50 (p. 210).
3. *Girvanella* mass in northeast corner and coatings of filamentous algae surrounding sections and fragments of Dasycladaceae, x30 (p. 210).
4. *Mizzia minuta* Johnson and Dorr. Longitudinal section showing outline of rounded protuberances of "pores" and character of wall. Transverse section of same species, x30 (p. 205).
5. *Anthracoporella* sp.? Transverse section of a fragment, x25 (p. 209).

FIGURE 1

FIGURE 2

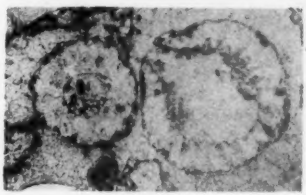

FIGURE 4

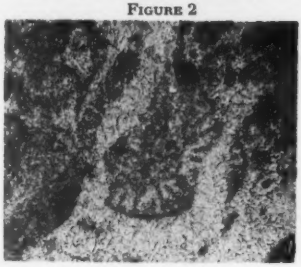

FIGURE 5

FIGURE 3

*GIRVANELLA, MIZZIA, ANTHRACOPORELLA*
Calcareous Algae, Guadalupe Mountains.

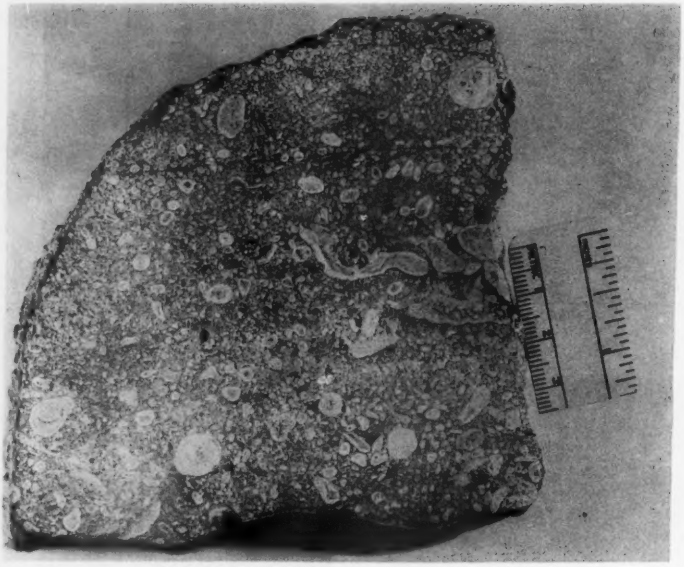

FIGURE 1

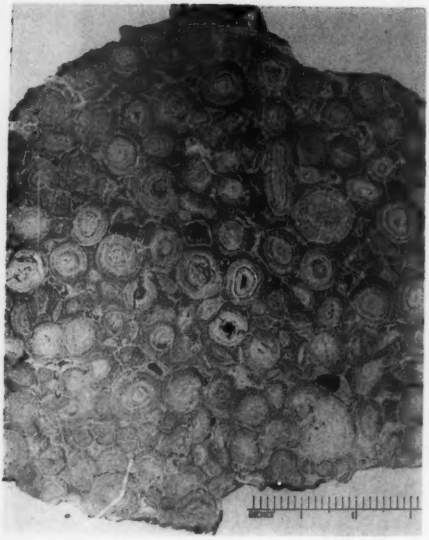

FIGURE 2

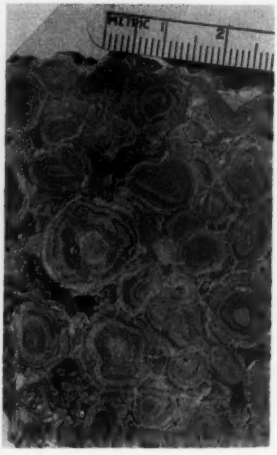

FIGURE 3

ALGAL "PISOLITES" AND CRUSTS

PLATE 6.—*Algal "pisolites" and crusts*

Figure

1. Algal "pisolites" and crusts. Specimen sawn and etched, x0.8. Walnut Canyon, New Mexico. Note transitional forms between the crusts and the "pisolites" (p. 213, 214).
2. Algal "pisolites." Sawn specimen, x0.6. Walnut Canyon, New Mexico. Shows irregularly developed growth layers of a typical pisolite (p. 213).
3. Algal "pisolites." Polished specimen, x0.9. Rattlesnake Canyon (p. 213).

PLATE 7.—*Mizzia, Collenella, Diplopora, Macroporella*

Figure

1. Undetermined Dasycladaceae fragments, x30 (p. 201).
2. *Mizzia yabei* (Karpinsky), x50 (p. 207).
3. *Collenella guadalupensis* Johnson, n. gen., n. sp. Hand specimen, x0.75. Dark Canyon, New Mexico (p. 212).
4. *Diplopora* ? sp. or *Macroporella* sp.? Longitudinal section of a large specimen, x35. Near Whites' City (p. 209).

FIGURE 1

FIGURE 2

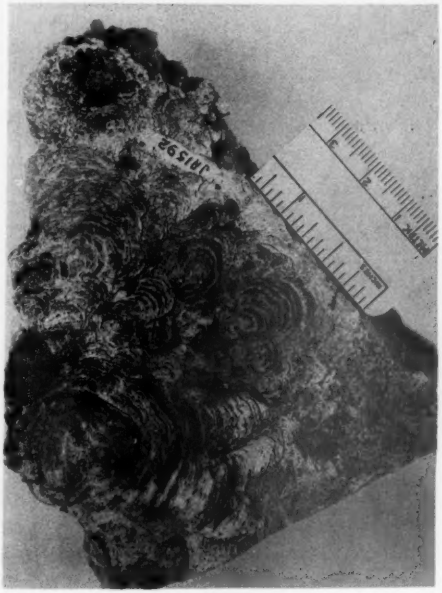

FIGURE 3

FIGURE 4

*MIZZIA, COLLENELLA, DIPLOPORA, MACROPELLA*
Calcareous algae, Guadalupe Mountains.

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# TILTED POSTGLACIAL LAKE BEDS IN OHIO

BY GEORGE D. HUBBARD AND RUTH C. ROCKWOOD

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## ABSTRACT

Abandoned lake beds in Ohio have been studied and mapped. A dozen have been examined, and five or six other areas have been studied; no evidence of postglacial lake occupation has been found.

The best evidence of the former presence of a lake is in the silts. Occasional traces of shore lines, four deltas, and a few examples of outlet channels have been found.

Tilting is shown by the distribution of the silts. They have been mapped up the slopes as high as they can be found. Erosion is recognized as a modifier of this distribution. Deltas and outlets identify tilting.

The direction of tilt is consistently nearly southward. The amount of tilting varies from about 2½ feet to 4 feet per mile.

Care has been taken not to mistake moraine and outwash terraces for shore lines. Both are recognized in valleys of central Ohio.

## INTRODUCTION

### GENERAL STATEMENT

The lake beds here described are all early postglacial; in fact two of them actually had ice contacts. No doubt any interglacial beds have been obscured so that there is probably no confusion between them and postglacial beds. Perhaps water, gas, and oil wells drilled in some of the old valleys have penetrated beds of clays and silts belonging to lake beds of interglacial times. At least one example occurs in the Killbuck valley where the wells of the Millersburg waterworks penetrated about 120 feet of fine white to gray sand and silt called by drillers "lake sand." This sand underlies 12 feet of flood-plain silts and sands and 14 feet of Illinoian gravel whose lower layers are cemented; it overlies 5 to 6 feet of water-bearing gravels reported to contain green pebbles, lying upon

bedrock. The materials of this well section plus the outwash gravel and sand terraces of the valley represent a valley fill of at least 210 feet, but although the wells are near the center of the valley they may not measure the maximum depth, and the 120 feet of sands and silts in the well section may be deposits from an interglacial lake.

The postglacial lake beds are distributed across Ohio (Fig. 1) from the Celina or St. Mary's reservoir in Mercer, the border county on the west, to Pymatuning reservoir on the Ohio-Pennsylvania boundary. Even Portage, the only county in a row across the State that does not contain parts of one or more of these dry lake beds, at the present time has many small lakes.

Some of the lake beds are too small to warrant study. Several areas suggestive of lake beds were studied but have no records of continuous lake occupation; an example is in a north-south valley east of Nelson's ledges, where very considerable lake sediments were deposited in several small ponds. The Killbuck valley (Hubbard, 1908, p. 239) was once supposed to have held a long finger lake, but the present studies have proved that the valley contained no lake in postglacial time.

The relation of these lake beds to Wisconsin moraines (Fig. 2) makes it clear that all are postglacial but that some of them formed while the ice was still near; hence some are justly called proglacial lakes.[1] Presumably they formed as the ice wasted and continued through the centuries until in each case the life history had been run, some ending long ago, some continuing as marshes down to the present.

## PREVIOUS WORK

In the early geological survey of Ohio the most persistent marsh and peat areas, notably the Willard, Scioto (Winchell, 1874, p. 352-354), and Orrville (Read, 1878, p. 529) sections, were recognized as lake beds but not as remnants of postglacial lakes. The county reports in *Geology*, volumes 1, 2, and 3 of the Ohio Geological Survey, mention swamp areas under several names, such as prairies, oak openings, peat bogs, and muck farms. Some workers recognized that peat bogs were late stages of former lakes.

Leverett mapped and published on the glacial features of Ohio giving more detailed information on moraines, old drainage lines, outwash deposits, and lakes than probably all other workers combined. His map (Leverett, 1902, Pl. II) shows only two lake beds other than those in the Great Lakes basin, the Willard and another in the northern part of Wyandot County between Carey and Sycamore. The Willard bed was

[1] Some of these were called proglacial on Ohio Academy of Science programs in 1938 and 1939, and again on the program of the Geological Society of America in December 1938 in New York.

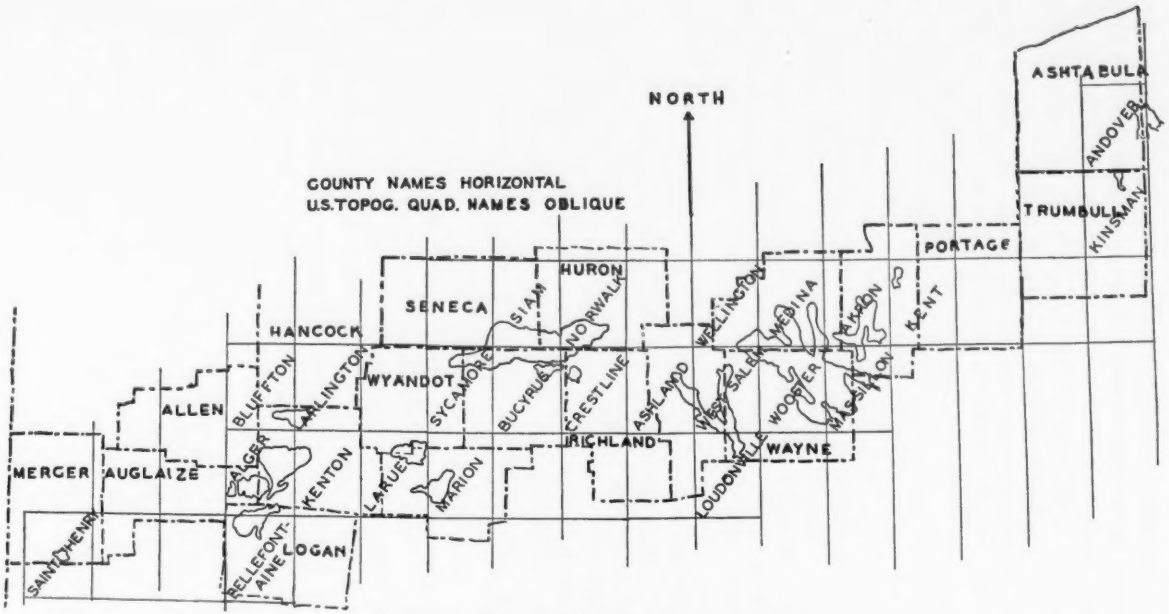


FIGURE 1.—*Map of the lake bed belt across Ohio*
Showing quadrangles, counties, and outlines of lake beds.

verified by the writers, but the second area was thoroughly traversed for silts in vain. Descriptions of local lake beds in his publication are very meager.

Dachnowski (1912) published results of his survey of peat bogs of

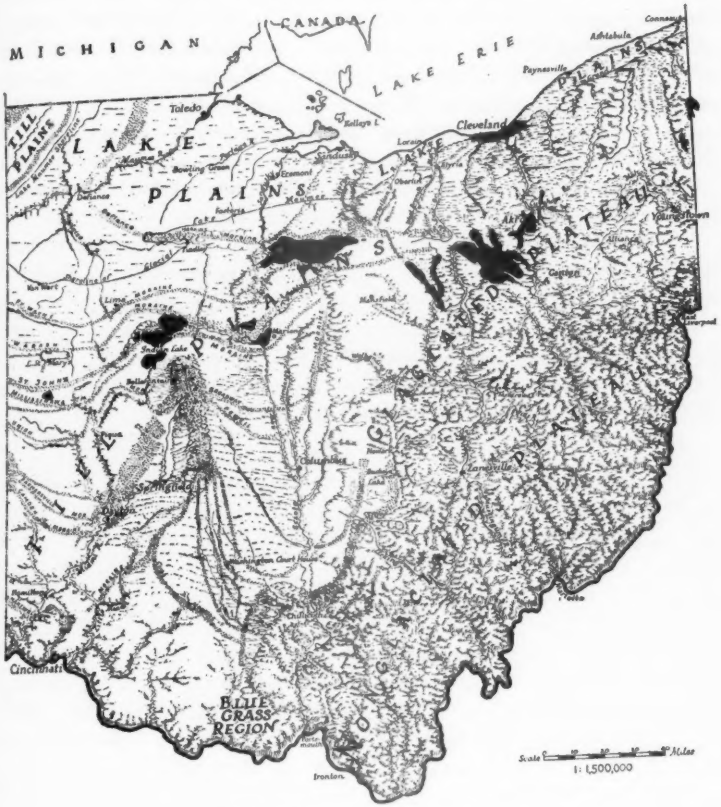


FIGURE 2.—*Physiographic diagram of Ohio*
Moraines in hachures; lake beds in solid black. After R. B. Frost.

Ohio. He added much to the physiographic knowledge of these areas, but as a botanist his problem was plant distributions and associations. He recognized all the lake beds that contain peat. Some of the mapping was good, some seems to have been careless, extending peat areas far out over moraine areas which no lakes ever occupied and omitting consider-

able areas of lake bed where apparently peat had been destroyed by oxidation.

Inasmuch as all the above publications antedated the U. S. Geological Survey topographic maps, greater accuracy was often impossible.

The U. S. Soil Survey made valuable contributions to the abandoned lake bed studies. A survey of Marion County (Conrey, 1916) revealed a lake bed west of Marion. Conrey (1921) recognized the Craigton lake bed, also the lake bed in the northern part of the county extending from Lodi past Rittman, and its extension southward to Orrville and Fox Lake. He recorded silts in many places included in these areas. In his surficial geology he correctly mapped alluvium along the Killbuck valley plains entirely across the county but, in another area near by, he did not distinguish between outwash and glacial lake deposits.

Before 1914 the senior author discovered that Lake Craigton's shore line was not level and strove to identify enough remnants of shoreline features, such as beaches, nips, terraces, and deltas, to make mapping and measurement of tilt possible. A wye-level was taken into the field, and from perhaps a dozen such remnants levels were run to government bench marks. Results, published in two papers (Hubbard, 1914a; 1914b) showed a tilt of about 70 feet in 18 miles.

### FIELD WORK

A total of 20 weeks in the summers of 1938 and 1939 were devoted to detailed studies of abandoned lake beds and areas suspected of being lake beds, including searches for outlets and ice contacts. The senior author has seen everything described here; the junior author participated in all the work in 1938 and a part of 1939, giving particular attention to geographic aspects of the work. John J. McKelvey worked 5 weeks in 1939, and other helpers shared the work in other weeks.

Oberlin College made appropriations for each summer from its research funds, and the Ohio Academy of Science twice granted assistance from its Emerson Macmillan Research fund. These grants were used both in field work and in preparing drawings and photographs.

Field methods consisted in reconnaissance work with a car, then detailed investigation with a spade, pick, and soil auger over every area. Every road which crosses a lake bed was traversed, and many foot traverses were made between roads, across fields, along streams, dredge ditches, and pipe lines. When a lake bed was found, observations on the sediments were made toward the margin in many radial lines to locate as accurately as possible the top of the silts, shoreline features, and delta or beach terraces. Nearly every sample handled in the field was scrutinized with a hand lens. The altitude to which silt

could be traced was determined by leveling to the nearest bench mark or altitude given at a road intersection. Scores of such determinations have been made. In places interpolations had to be made with reference to houses, contour lines, and other identifiable data. It is believed that altitudes are generally more accurately determined than are the silt margins and beach features.

Altitude determinations and other physical data were plotted on the government topographic sheets from which the maps were traced. Samples of critical material were tested by screening and by settling, decanting, and microscopic measurements. This last testing was done by Miss Jane Keeler in the research laboratories of the Department of Geology and Geography of Oberlin College under the direction of Dr. Fred Foreman.

ACKNOWLEDGMENTS

Since this study began many years ago under the State Geological Survey, thanks for opportunities and assistance are due Drs. Edward Orton and John A. Bownocker, past State Geologists, as well as Wilber Stout, the present State Geologist. Thanks are also due field assistants, always keen, willing, and resourceful; the hundreds of local citizens who individually and in groups have furnished information and friendly interest; members of the Ohio Experiment Station and Soils Survey staffs for instruments, maps, and counsel; the staff of the department of Geology and Geography of Oberlin College for counsel and for space and equipment granted for work; and Oberlin College and Ohio Academy of Science for financial grants. To Dr. Frank Leverett for his critical challenges and discussion, a special mention is due, for this stimulus has done much to make for clarity and completeness in the paper.

## PROBLEMS

This study of postglacial lake beds presents three groups of problems. The first is the geomorphology of an area molded in turn by ancient stream, glacier, lake, and postglacial stream. Valleys and basins were formed by preglacial and interglacial streams and by glacial erosion and deposition; then lakes built their beds and fashioned their shore lines; and finally came postlacustrine erosion and deposition and accumulation of wasting vegetation. What shall be charged to each agent and process? How can one discover the lake beds and map and interpret them long after the water has gone and the modifying agents have worked upon them?

It soon became obvious that some of the beds were not only modified by streams but were also disturbed by diastrophism. Thus appeared

the second group of problems whose solution has become the prime motive for this paper.

The third group of problems includes the economic resources and their utilization, changing as the decades have passed. This is reserved for another paper.

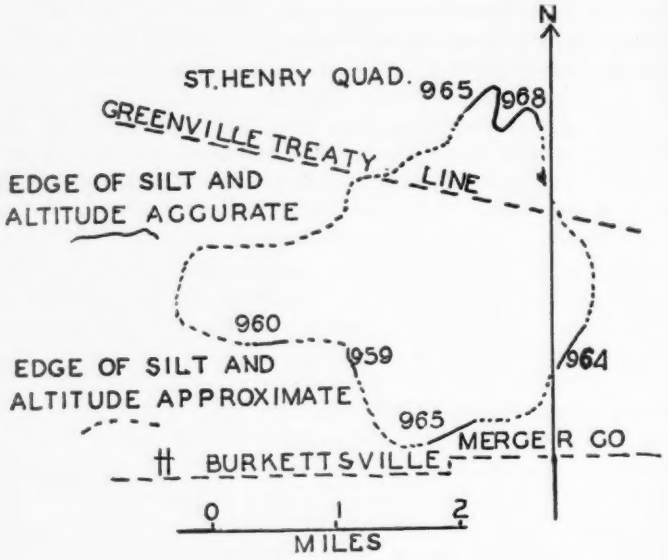


FIGURE 3.—*St. Henry lake bed*

## DIFFICULTIES

At first shoreline features were considered most important criteria, and no use was made of lake-bed silts. Two difficulties, however, disturbed from the start. It was impossible to find even approximately continuous shore lines, and some trouble was found in distinguishing between shoreline sands, gravels, and terraces on one hand and outwash or moraine terrace materials and forms on the other. The paucity and fragmentary condition of beach and shoreline materials and forms promoted intensive search until lake-bed silts and clays were discovered approximately up to the shore line. There is no danger of confusing bedded silts and varved clays with outwash or with kame and moraine terraces and ice-contact phenomena, even though shoreline materials might be so confused. Some valleys have moraine terraces and related features, others do not, and it is believed that they have been fairly successfully

differentiated. Several lake beds so located that they could have been occupied by ice tongues or drainage waters are now covered with the same sort of silts and clays as occur in certain valleys that must have been occupied by ice tongues and for melt-water drainage. In four places distinct deltas are known whose levels correspond with the lake bed silts.

Another difficulty arises from long-continued erosion. In places silts have been completely eroded from parts of lake bed areas lower than other near-by more level, but higher, areas still well covered with silt. Erosion from hillsides above lake beds has supplied waste which now covers lake silts and even peat areas. Thus in places the surface material is alluvium even though we have abundant evidence of lake deposits a few feet underneath. Cultivation has also mixed thin silt deposits with drift below or with alluvium above. Occasional lake beds in lateral valleys which were marginal to ice tongues contain silts too high to correlate with the main lake bed. To know that such difficulties exist is a valuable warning to the investigator.

## DESCRIPTION OF THE LAKE BEDS[2]

### MERCER COUNTY

The St. Henry bed 3 miles southeast of the town of St. Henry measures about 3 by 4 miles (Fig. 3). The St. John's moraine bounds it closely on the north, and the Mississinawa moraine is near on the south. The Wabash River in its loop into Ohio flows through the lake bed in a dredged ditch, as do three of its tributaries. Silts are found all over the area, clays in some parts, but no peat; no shoreline features were recognized. The edge of the silts was found on the south side at altitudes of 960, 959, 965, and 964 feet; in the north where they reach the moraine they are at 965 and 968 feet. If this were the only lake bed studied, the difference of about 4½ feet between the average of the opposite side altitudes would not be stressed, but this difference, both in direction and amount of inclination, is in full harmony with the rest of the beds studied. Hence this bed is counted as one that shows tilting.

### ST. MARY'S AND LORAMIE RESERVOIRS

St. Mary's, Grand, or Celina reservoir a little farther north in Mercer County and between the St. John's and Wabash moraines covers all the lake bed which was drowned 100 years ago to make this canal reservoir. A careful search entirely around the present lake revealed no

[2] The lake beds are described in order from west to east.

lake clays or silts and no abandoned shore features. The drowned lake bed deposits are completely covered.

Loramie reservoir seems to occupy a portion of a stream valley and not a lake bed. In the Ohio Geological Survey County report Hussey

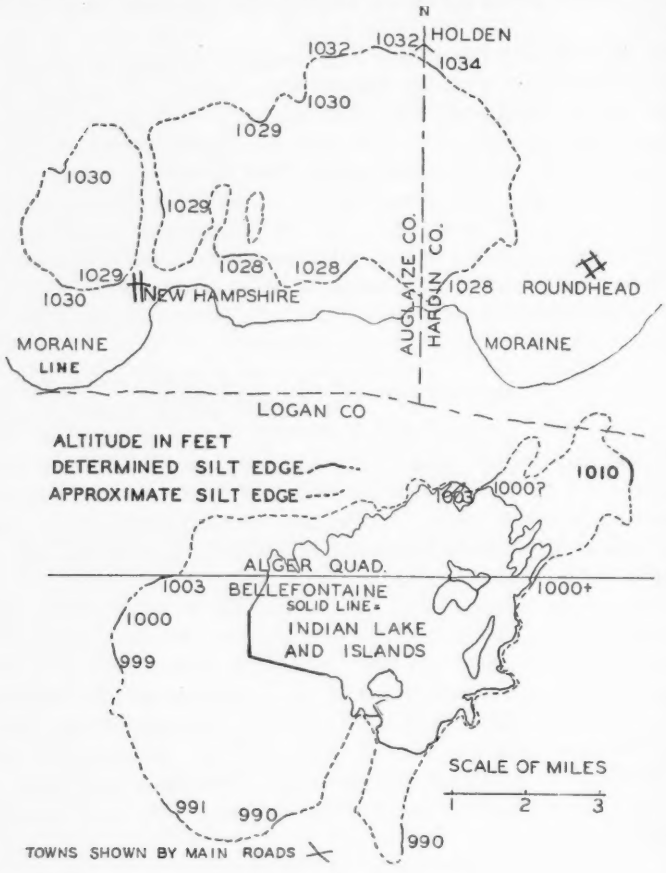


FIGURE 4.—*Indian Lake bed*

(1878, p. 457) says that forests were incompletely cleared from the area to prepare for the reservoir, a statement which suggests a valley in till plain rather than a lake bed. No silts or clays were found above present water level.

LEWISTOWN RESERVOIR OR INDIAN LAKE

The Mississinawa and St. John's moraines cross this area and separate a northern abandoned lake bed, Holden, from a southern one now nearly covered by Indian Lake. When the Lewistown reservoir was first made, it drowned three small lakes, remnants of a single larger lake. Peat areas reach beyond the present shore lines in the north and east, and silt areas extend a short distance northeast, but the large area of silt lies to the southwest. The long bulkhead prevents the water from

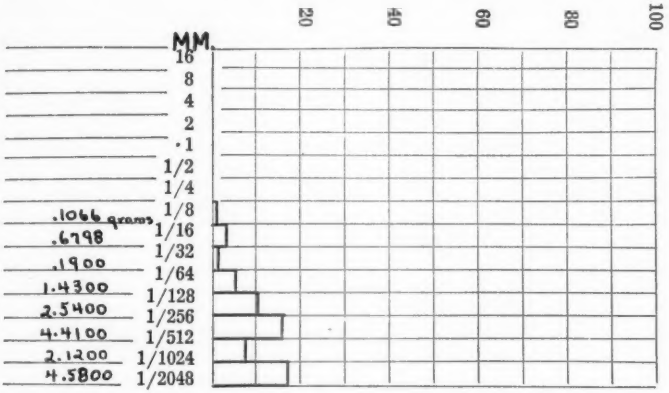


FIGURE 5.—*Histogram of varved sediments*

flooding these silts which extend back 2 miles from the present lake. The water level today is 997 feet. The map (Fig. 4) shows a NE.-SW. inclination of about 20 feet in 7 miles. Silts are over 3 feet thick near the northeast edge of the lake bed. Just west of the lake near Lakeview a drilling over 7 feet deep revealed 4 feet of varved sediments at the bottom under clays and silts. The histogram (Fig. 5) shows the nature of the sorting in these sediments. This sample showed 24.63 per cent of carbonate. More varves were found a mile farther south. The tough clay in many drillings southwest of the lake and even in plow furrows is called "jackwax" by the farmers.

Across the north side of Indian Lake a narrow moraine not shown on the map connects eastward with massive moraine at Roundhead and northwestward with characteristic Mississinawa moraine.

Two or three miles north of Indian Lake another small member of the Mississinawa moraine connects eastward with the same Roundhead mass. It separates the Holden lake bed (Fig. 4) to the north from a plain to the south which, because it is free from silts and clays, is not

believed to be a lake bed. The map shows an inclination of 5 to 6 feet in 3 to 4 miles.

In the broader eastern part of Holden lake bed a level tract 2 miles east and west by 3 miles north and south, drained by open ditches and tile, is composed of rich black earth nowhere peaty. This tract has long

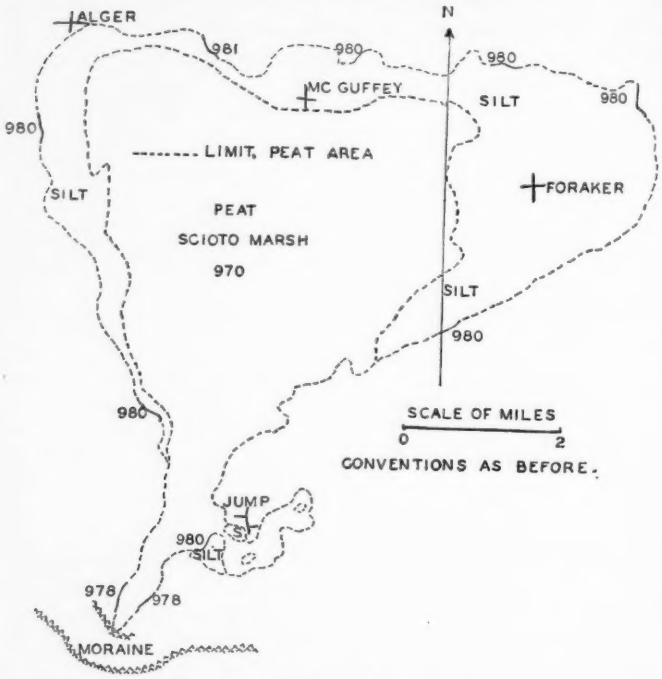


FIGURE 6.—*Scioto or Kenton marsh*

been called the prairies. Farmers live along its margin but not upon it. No shoreline features were noted around these two lake beds of the Indian Lake area.

### KENTON OR SCIOTO MARSH AREA

This lake bed has one of the largest peat areas in Ohio (Fig. 6). The peat area tapers southward for 5 miles, and the peat, often 6 to 10 feet thick, disappears between striking Mississinawa moraine areas. The south end of the valley was once closed by moraine where the Kenton road now crosses, but the Scioto River, both a feeder and the outlet for the lake, came down from the Holden lake bed and, gathering tributaries

near Roundhead, cut through the moraine. North of the moraine the channel needs dredging across the marsh and almost to Foraker, a distance of 8 miles.

Figure 6 shows the distribution of silt. On the east side a narrowing strip of silt disappears southward beneath the peat. It extends a mile or more farther south on the west side. Near Jump is a small patch of silt, but for nearly 7 miles around the tapering southern extension

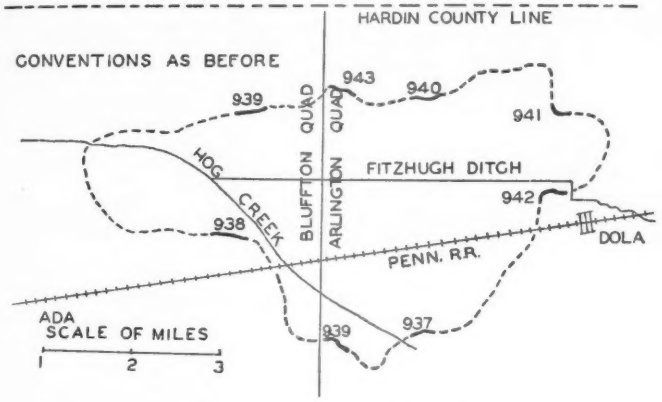


FIGURE 7.—*Hog Creek lake bed*

of the peat no silt shows beyond the peat margin. Shell marl is found below the peat in several drillings.

The altitude of the edges of peat in the southern part of the area is about 978 to 980 feet wherever determined except in an isolated embayment southeast of Jump where it is almost 990. In the north, silts rise beyond the peat very consistently up to 980 feet. It would be difficult to discover the edge and altitude of the silt in the south where covered with peat, in places more than 10 feet thick, but, inasmuch as peat does cover all silt and yet is not above 980, the silt margin in the south must be lower than in the north, indicating a little inclination up in the north. The distribution of the peat suggests that the bed was disturbed before it was drained and that the peat developed shortly after.

### HOG CREEK LAKE BED

In northwestern Hardin County (Winchell, 1874, p. 352-354) lies the next lake bed. Hog Creek, which now drains this area, drained the lake by flowing over thickened drift. The creek across this drift obstruction

2 or 3 miles west of the lake and about 2 miles north of Ada now has a gorge about 20 feet deep. The Ft. Wayne moraine is near on the north.

The lake bed is about 6 miles east and west by 3 miles north and south (Fig. 7) and is largely covered by black earth, peat, and peaty soils. Silts entirely girdle the black earth area. In a number of places where black earth was penetrated by boring it was underlain by silt. Nearly 2 miles northwest of Dola on the east-west road a 10½+ foot bore showed

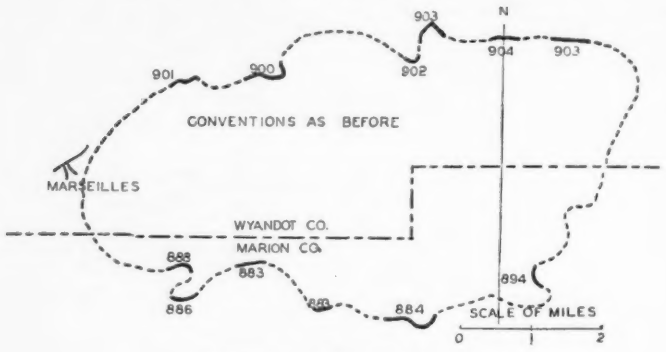


FIGURE 8.—*Marseilles lake bed*

1½ feet of peaty black soil and 9 feet of blue-gray, poorly bedded silt. This bore was a full auger length, and the bottom of the silt was not found.

Average altitudes recorded on the map show an inclination of 3 to 4 feet in 4 miles in a NNE. direction. The black earth part of the lake bed is so level that all drainage is artificial but so effective that good crops of beets, corn, barley, and alfalfa are grown.

### MARSEILLES LAKE BED

The authors discovered this bed 2 or 3 miles east of Marseilles and followed its deposits out to the margins. The auger hole made at the first point disclosed jackwax 3 to 4 feet thick. It is reported even thicker in deep post holes. Usually silt in the central part is covered by this dense clay and, in Pitt township (Winchell, 1873, p. 626) 2 to 3 miles east, by peaty soils though no true peat is known.

Silt is better exposed on the north side of the clay than on the south, perhaps because the actual south side is not preserved. Prairie Run and one of its branches, Thompson Ditch, flow along the south side. No silts or clays were found south of the streams or north within a mile of them on some cross roads. As the bed now stands, rising to the north, there is no barrier north of these streams, but if the tilting were

restored, thus lowering the north side 15 to 20 feet, the central part would be lower than the south side. Silt determinations recorded on the map (Fig. 8) show a northeast inclination of 20 to 21 feet in 6 miles. Toward the extreme eastern end of the north side, ¾ mile south of Harpster, the slope is steeper than elsewhere, from 912, 8 to 10 feet above where the beach should be, down to 889, and all silts ever laid on it are gone; east of the road they are cut away by tribuaries of the

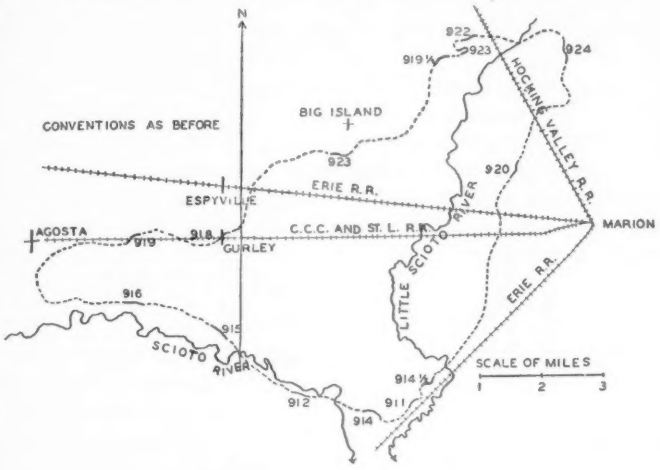


FIGURE 9.—*Marion lake bed*

Little Sandusky River but west of the road they are preserved up to 902 to 904 feet.

The small streams that drain the bed flow southward across it, and the tributary of Tymochtee Creek that receives them follows closely the south side, which condition suggests that the tilt occurred while water was still present.

Since no moraine borders this bed on the north, neither ice nor moraine retained the waters there, and the present north-flowing streams, Tymochtee and Sandusky, soon established themselves; but it is believed that at first the waters were obliged to work their way westward and southward.

### MARION AREA

The soil map of Marion County (Conrey, 1916) revealed this lake bed. The shore line was first found 4 miles southwest of the railroad center of Marion (Fig. 9), where the silt and sandy gravel edge of

deposition and the little nip and terrace of shoreline forms were accurately located by leveling from the bench mark (921) at the crossroad. The water surface was placed at 914½ or 915 feet, correct within 6 inches.

The figures on the map show the altitude of the silts at areas on all sides of this lake bed. A definite moraine barrier crossing the valley

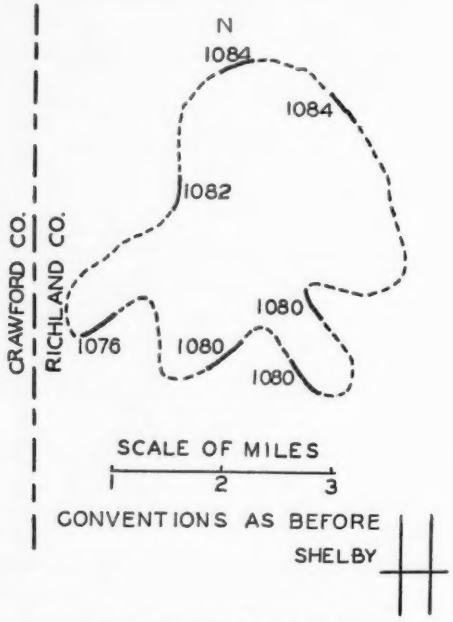


FIGURE 10.—*Shelby lake bed*

where Rock Fork enters the Little Scioto gives present silt at altitudes of over 920 except where the streams have cut it away. North of it there may be some lake deposits, but if so they are higher than those to the south; they rise much more rapidly northward than any lake beds studied and they resemble flood plain or are mixed with flood-plain materials much more than are ordinary lake-bed deposits. Probably no lake existed north of this drift barrier at Rock Fork mouth. No peat was found in this Marion lake bed, but areas of jackwax occur, especially south of Big Island 4 miles west of the Marion railroad junction. The map shows an inclination of the surface of 10 feet in 7 miles.

Actual shore features were found in two places only, and nothing resembling outwash or moraine terraces.

### TABOR SCHOOL AND SHELBY AREAS

These two areas are little more than large irregular kettles (Fig. 11). The silts are typical of lake beds, attain thicknesses up to 3 feet, and in a few patches are still overlain by peaty soils, muck, or black earth. No shoreline features were found or expected on such small and irregular lake beds, and because of their size the lake beds show little systematic differences in altitude of silt borders.

That these are lake beds, however, is shown by the silts, peat, crops, levelness, and dredge ditches. The first is mapped solid moraine by Leverett (1902, Pl. XIII), but the moraine distinctly divides into two ridges 4 miles or more apart across the Bucyrus quadrangle. The second, Shelby (Fig. 10), lies south of the moraine after the Ft. Wayne and Wabash with their divisions merge into the interlobate mass between Scioto and Black River lobes. Shelby bed shows an inclination of surface of 4 feet or more in 2½ miles.

### WILLARD-CELERYVILLE AREA

The Willard-Celeryville lake bed, one of the three largest, occupies parts of six topographic quadrangles (Figs. 1, 11), lies against the Defiance moraine across most of its northern side, and is part of what Leverett's map (1902, Pl. XIII) shows either as a drainage channel or an arm of the great proglacial lake in the Maumee stage. This lake bed is nearly 200 feet higher than the Maumee shore line at Findlay, where this "channelway" reaches the Maumee level. It does not seem possible to consider this area as either a part of an active drainage channel or an arm of Lake Maumee.

Since the Defiance moraine along the northern margin of the lake is almost continuously lower than the high altitudes of silts on the south side, it must have been assisted by ice in retaining the water in its early high stages. At this stage the moraine and presumably the ice wall (end of the glacier) swung far enough south in Wyandot County to meet the Ft. Wayne moraine, there 1000 feet high, and thus hold more or less steadily for a short time the Willard Lake at its high level. As the ice withdrew from northern Wyandot County to its longer stand to build the Defiance moraine in Seneca County, the lake waters fell as low as possible with the erosion taking place by the tributaries of Sandusky River. While the ice was building the Defiance moraine, the river is believed to have delivered its waters westward, as from the start, through a valley now containing The Outlet. But

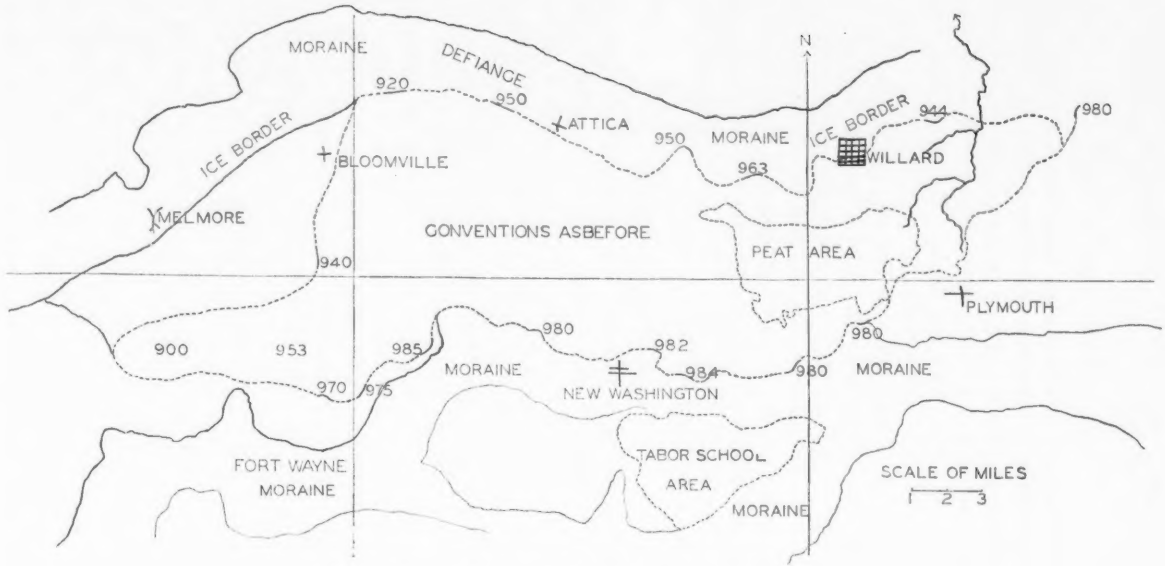


FIGURE 11.—*Willard lake bed and Tabor School lake bed*

when the ice melted back from the moraine, the Sandusky River found a lower course across the drift to the successively lower levels of Lake Erie. During later stages of the ice withdrawal, Honey Creek carried more water and made a larger channel than it needs today; but later tributaries of Huron River offered lower outlets and shared the drainage of the Willard Lake area as today, and Honey Creek was reduced about to its present size.

This digression from the lake bed to its drainage is made because the complicated history of the latter has never been described. In the west some of the most desirable records of these changes are already eroded away.

The peat area covers at present 12 to 13 square miles, while the silt area is roughly 18 by 5 miles or nearly seven times as large. The extremely level peat area (Pl. 1, fig. 1) is underlain by silts and clays. Evidence in vegetation suggests that perhaps half the silt area was once covered by peat. The peat area is becoming smaller every year by the oxidation of the organic matter, tremendously augmented by drainage, cultivation, and occasional fires.

This lake bed is narrow from north to south and could not show great inclination at best, but, with the north shore on the ice border, one can make no comparisons for unlevelness. When the water surface fell to the levels longer occupied, a new water margin on the south was established on silts already there, but their presence makes it difficult, if not impossible, to locate the new south shore line that was made when the present north shore line was recorded. The usual silt distribution method is useless here, nor is the peat margin of any value, for the peat has wasted much; its growth comes down to recent time, and its margins are about the same height all round. Shoreline features are scarce. For these reasons this lake bed is not used in the studies of tilting.

### CRAIGTON LAKE AREA

In the lake bed named Craigton extending from Custaloga and Big Prairie at the south to Ashland and Nankin in the northwest and to Redhaw and Pleasant Home in the north are preserved the evidences of the longest tilt, about 18 miles, though not the greatest per mile, and the first evidence of tilt in Ohio found away from the Great Lakes. The senior author first saw this lake bed near Funk in 1907 (Hubbard, 1908). Later it was more fully explored and described briefly as a Finger Lake. Not until 1913 was the inclination of its surface discovered and a more careful survey of altitudes made. The effort then was to establish shoreline and delta levels; the value of silts as a criterion for the presence of a lake was not yet appreciated. While it is probably true

that bits of shoreline evidence were correctly used (Hubbard, 1914a; 1914b) the conclusions were properly challenged by Leverett (personal correspondence) who had used this valley as presenting an example of

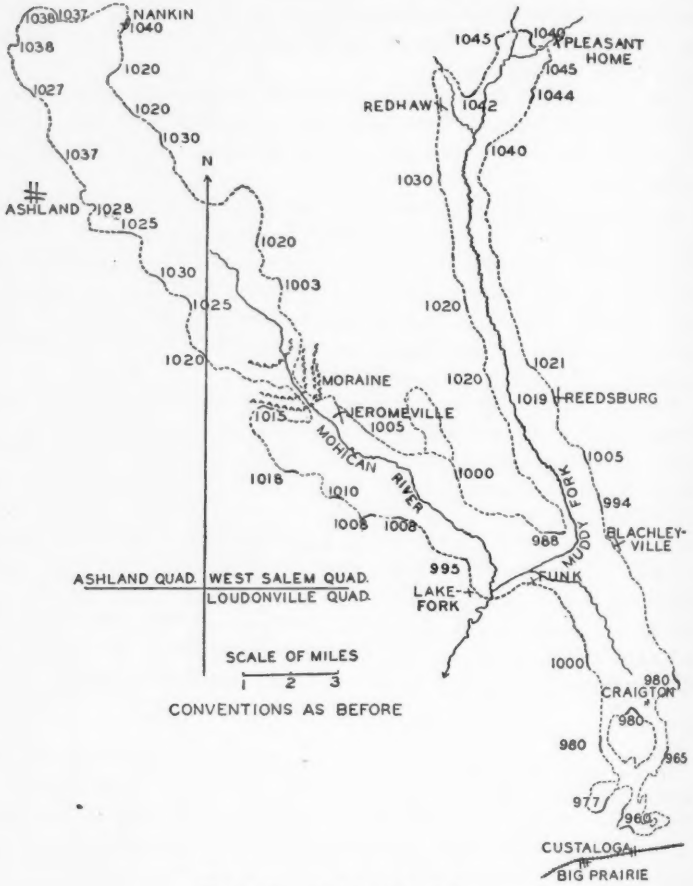


FIGURE 12.—*Craigton lake bed*

moraine terraces and remnants of marginal ice-drainage deposition. This difficulty was removed when the mapping was done by silts across the lake bed, deposited indiscriminately over moraine, outwash, and other valley filling, and not by remnants of shoreline features.

The outlet for this whole lake bed is now southward through Lake Fork of Mohican River (Fig. 12). The stream takes the water from the two lake arms at their junction a mile west of Funk to make Lake Fork. Apparently this stream has been the outlet for these valleys since the last ice withdrew, except for a relatively short period when the ice still lay far south of this outlet leaving room at the south end for only a tiny lake which drained out among moraine and stagnant ice blocks. Even now ditches gather a little water from several marsh tracts in and near the south end of the lake bed and carry it 2 miles to Odel Lake whence it too flows to Lake Fork. Leverett has pointed out terraces at 980 feet 2 miles south of the village of Lake Fork on Lake Fork, and moraine in the valley was cut through in the early years of the flow from Craigton westward and down Lake Fork. If drainage were to start in these valleys with tilt restored and ice out, but with the drift uncut in the present west valley, the water would certainly drain to the south. Therefore, the first deepening of the present course of Lake Fork as it flowed from the town of Lake Fork to the master valley, in which are Shreve and Loudonville, must have begun while ice occupied the lake bed, and at that time so lowered the channel through drift and outwash that, even though the lake bed suffered a notable tilt, the drainage continued to flow out by the Lake Fork route.

A strip of peat areas about half a mile in width extends south of the outlet almost to Custaloga. Silts are found over all the lake bed but they are covered locally by slope wash and by fans along the margins and where streams come into the plain. Moraines partly fill the valley just below Craigton, but the main obstruction in the valley here is a rock hill ¾ mile across. In the west arm midway between Funk and Ashland a moraine loop swings across the valley. Three or four distinct ice fronts occur in this mass and, while well breached at present, they may at some time have constituted a barrier across the valley. There seems to be no way to prove that they did because the silts are essentially the same height on opposite sides of the moraine remnants, and the silt deposits through the valley are consistent in height, rising to the north.

In the upper part of this western arm from Ashland north are many subdued moraine forms which have silt over their flanks and summits as if the lake had washed over all. None are now higher than the silt levels on the valley walls, though some of the more level-topped ones probably once rose above the water and were leveled down.

At the upper end of each lake arm are two deltas, one at each "corner" where present streams come in. At Pleasant Home (Pl. 1, figs. 2, 3) the

delta attains altitudes of 1040 to 1045 feet, consists of delta-bedded sands and gravels, and is surrounded on the lake bed and underlain near its edge by clays, stiff and pure enough for tile and brick-making, which were sorted from other stream-borne waste during deposition. At Redhaw (Pl. 2, fig. 1) the delta surface is more intact and by leveling is found to be at 1042 feet. At Nankin the altitude is 1040 feet. The west-facing front of the delta and its unscarred outer lip are well marked near the edge of town (Pl. 2, fig. 2). Nearly 2 miles west of Nankin is the fourth delta whose elevation is 1037 and 1038 feet (Pl. 2, fig. 3). It has a remarkably well-preserved front and outer lip.

All four deltas have been extensively opened for gravel and sand. Thousands of loads have been hauled for road and other construction, but enough still remains so that the delta form and structure are clear. No delta fronts are opened.

Edges of silt—limits of lake bed—have been marked in more than 40 places around this lake bed. The edge does not seem to rise uniformly, but probably the determinations of the true edge are not perfect (Fig. 12). If one takes averages of three or four readings in a vicinity he probably has a truer value than if he takes the lowest readings. The higher levels in any locality are probably truer than the lower ones because silts could not be deposited above the water but could be lowered by erosion.

East of Jeromeville a small valley leading back between the two arms and drained by Glenn Run has silts for a mile or two up to altitudes of 1035 and 1040 feet. No silts in this valley were found at 1050 and 1060 feet. A little mass of drift at the entrance partly plugs the valley there but leaves it wide and open farther north. A tongue of ice in the main valley probably ponded waters in the lateral valley long enough for the silts to be deposited before the main lake was formed.

Large gravel deposits have been opened near Funk, better sorted and more abundant than in any delta. These piles of gravel were interpreted as kames made at a notch between the two tongues of ice that occupied the respective valleys on the two sides of the deposit. They are not delta-bedded nor in a favorable place for a delta.

The inclination in this lake bed when taken from deltas and remnants of shore lines over 25 years ago was placed at 70 feet in 18 miles. The more recent detailed measurements and determinations give 80 feet on the east arm in about 18 miles, and a difference of 5 or 6 feet between the north and the northwest ends suggesting an east-west as well as a north-south component to the uplift, or a direction just a few

FIGURE 1. SURFACE OF WILLARD LAKE BED

FIGURE 2. DELTA EXCAVATION HALF A MILE WEST OF PLEASANT HOME

FIGURE 3. DELTA TOP NEAR PLEASANT HOME

LAKE BEDS AND DELTAS

FIGURE 1. DELTA EXCAVATED HALF A MILE NORTHEAST OF REDHAW

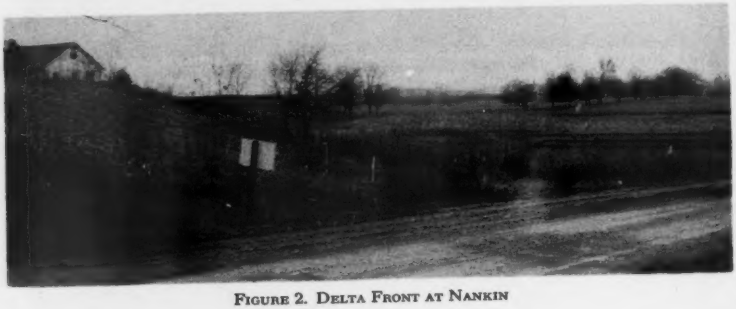

FIGURE 2. DELTA FRONT AT NANKIN

FIGURE 3. DELTA FRONT 2 MILES WEST OF NANKIN

degrees east of north for the maximum tilt. Most of the lake beds examined agree with this interpretation.

### MEDINA-ORRVILLE AREA

This area presents more problems than it solves. Earliest mention of lake beds here is in the old Ohio Survey reports (Read, 1878, p. 529; Wheat, 1878, p. 362-380). When the senior author first found this area three parts were described as having been occupied by Finger lakes (Hubbard, 1908). The Chippewa Lake valley was described as still possessing a remnant lake whose antecedent was supposed to have stood at about 1010 feet. The east-west valley from Lodi past Creston, Sterling, and Easton was recognized as another whose waters were believed to have stood at a little less than 1000 feet. A lesser valley, called the Orrville basin, was thought to have once been filled to about 1020 feet.

In the summer of 1938 silts were discovered at 1040, 1050, and in places above 1100 feet. It was found that the boundaries between these three Finger lakes were sufficiently flooded so that they coalesced, and the River Styx valley also contained an arm of the enlarged lake. Instead of three or four Finger lakes there was perhaps one "whole hand" lake with seven or eight digits, and these not very cleverly arranged for a hand (Fig. 13). The highest silts in this area were in the south and not in the northeast as they had been in all others. The histogram (Fig. 14) shows the good sorting at silt-fineness taken at 1102 feet altitude and at bottom of a 3-foot bore. Another troublesome item was the fact that there are five outlets at or below the 1000-foot level today. One of these, the Tuscarawas River, flows through deep cuts in drift and rock from 1100 feet, or a trifle over, down to the present level. Another is an abandoned channel among moraine hills starting north of Canal Fulton and joining the Tuscarawas at Crystal Spring through a tributary called Mudbrook Creek. The third outlet is through Killbuck Creek leaving the area at Burbank over what seems to have been a moraine plug in a submature to young valley, probably interglacial.

The fourth, Newman Creek valley, is a youthful postglacial gorge from about the 1100-foot contour line down to present levels, and no doubt its stream once helped to drain water from the Orrville portion of the lake. It may have competed for some time with Sugar Creek, but the swell at Orrville in the lake bed is now a well-marked divide consisting of silt-covered till—subdued moraine.

The fifth outlet is down Black River fed by three streams which now flow into the lake area, unite near Lodi, and then flow toward Lake Erie. Silt in the vicinity of the junction of these three streams attains only

1040 to 1050 feet, and the preglacial valley only slightly obstructed by drift was wide open above the 1000-foot line. Obviously something plugged Black River valley during the lake stage that coincided with the deposition of the high silts. The Defiance moraine is strong west of the Lodi marshes and this Black River outlet and continues westward

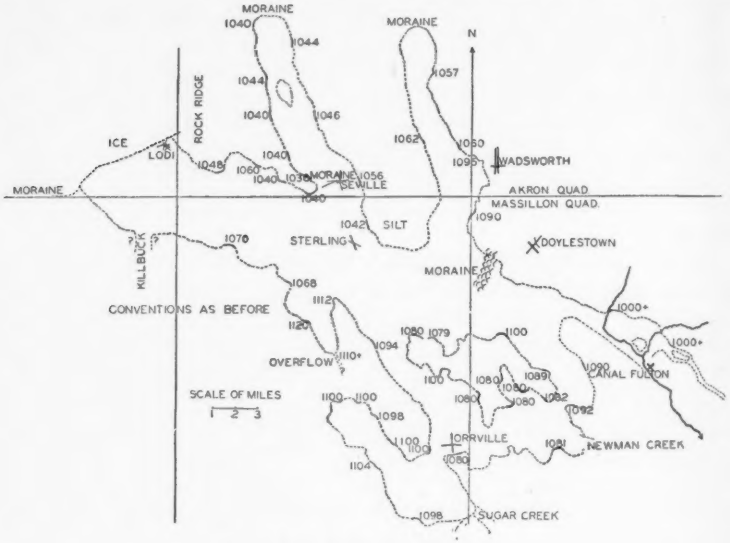


FIGURE 13.—*Medina-Orrville area*

for miles. A rock ridge of the Allegheny Plateau also extends from a point a mile or so east of Lodi several miles to the north with minimum altitudes of more than 1100 feet. Moraine runs from its northern part eastward across the Chippewa valley above Whittlesey, and across River Styx north of Boneta. Therefore, the ice was doubtless the barrier across Black River here while it built the Defiance moraine and these connections north of the rock ridge.

Several strong loop moraines cross Tuscarawas River valley between Crystal Spring and Clinton, another at Doylestown Station and Easton. One swings across the Styx valley a mile or more north of Rittman, a second a mile above the village River Styx, and a third, less of a loop, near Medina is probably a part of the Defiance moraine. In the Chippewa Lake valley a loop crosses at Seville and a more straightened moraine above Whittlesey which was mentioned above. The moraines

at Seville and above Rittman are probably of the same age but earlier than the Defiance; the ones in these same valleys near Medina are the same age as the one across Black River below Lodi from where westward it is designated Defiance Moraine. It is thus clear that a valley dependency of the retreating Wisconsin ice lobe lay in the great east and west valley of the lake bed. As it melted back, it built a series of moraines which effectively plugged what is now Tuscarawas valley past Canal Fulton; after more ice melting, the rock hill between Chippewa and Styx valleys showed through the ice, the two short dependencies temporarily occupied the valleys. As these tongues shortened, the lobe became free of dependencies in this region and lay with its margin across the Black, Chippewa, and Styx valleys long enough to build the great Defiance Moraine.

The distribution of these ice tongues has had much to do with the distribution of lakes and silts. When the large tongue reached below Canal Fulton, it must have ponded waters into the Orrville valleys setting water back into the area south of Orrville to the 1100-foot level, high enough to overflow down Newman and Sugar creeks. At the same time it was also set back northwestward to Smithville, but with the same outlets, and spread over all Orrville and its peat areas. It was then spread as continuous water from the ice southward through the Fox Lake area, the Orrville valley, and the Chippewa Creek valley from Orrville toward Sterling. The little valley whose stream starts two miles north of Smithville and flows northward must have been flooded to a higher level than the others. Silts here range from 1110 to 1120 feet, and the outlet over to the Smithville valley is now just above 1100. Ice pushed close to the shoulders of the rock hills at the north-facing mouth of this little valley and forced a little "marjelen-like" sea to overflow southward (Fig. 13). The Styx and Chippewa valleys were probably ice-filled while these valleys south of the ice tongue had initial lakes.

When the ice had withdrawn beyond Lodi and halted to build the Defiance moraine, the lowest outlet was toward the east down the Tuscarawas or the abandoned channel north of it, probably at first through both courses (Fig. 13). Killbuck also came into service soon after the ice passed Lodi. The water level at this stage must have been about 1040 to 1045 feet and fell to 1000 before the little, eastward, abandoned outlet ceased to function. Killbuck and Tuscarawas competed only until low divides near Creston and Easton appeared above the falling water. When the melting of the ice opened East Branch of Black River, Killbuck and Black contested for the Lodi marsh area, a contest not yet decided.

In many places, particularly on the upper slopes and hills, siltlike materials were found which were ascribed to weathering. They seem to represent horizon A of the normal soil profile. The matter of distinguishing between lake silts and this horizon of the soil profile was often discussed in the field. The silts shown in the histogram (Fig. 14) could

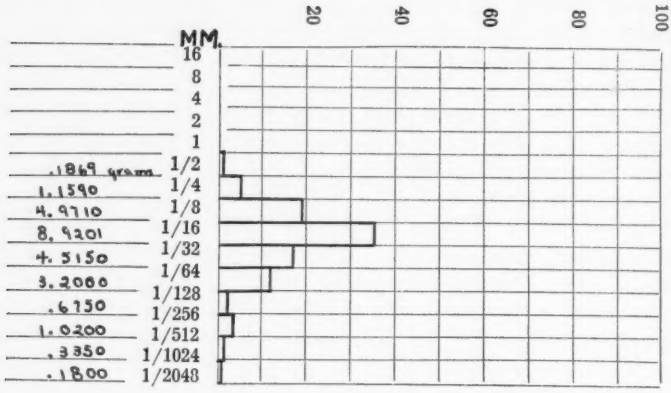


FIGURE 14.—*Histogram of silts near Orrville*

not be a part of a soil profile and be so well sorted at a depth of 3 feet.

Perhaps one could interpret the silt distributions to mean a tilt up in the south over a once continuous lake in all these silted valleys, but this is consistent neither with the moraines and recognized ice-tongue distributions nor with the inclination of the Craigton lake bed only a few miles to the west.

And so the comments on this Chippewa-Orrville area must close where they began, with abundant evidence of a great sprawling lake bed and several outlets, used at different times in a system of valleys occupied at times by active valley dependencies and later by stagnant ice blocks, but with no positive contribution to the problem of postglacial tilting in Ohio.

### AKRON AREA

About 3 miles west of Akron lies a valley in places marshy, in other places covered by small lakes, and in still others supporting industry as at Barberton and New Portage. It has been interpreted as a part of an overflow channel (Leverett, 1902, Pl. XIII) from melting ice in the north to open drainage lines and the Tuscarawas River in the south. Outwash gravel and sands beneath silt in Barberton excavations substantiate the claim. It was certainly earlier occupied by an ice tongue

or valley dependency, which built several loop and lateral moraines such as that one a mile south of the Summit-Wayne county line. Moraine terrace was found a mile north of Shocalog Lake (Fig. 15) above lake

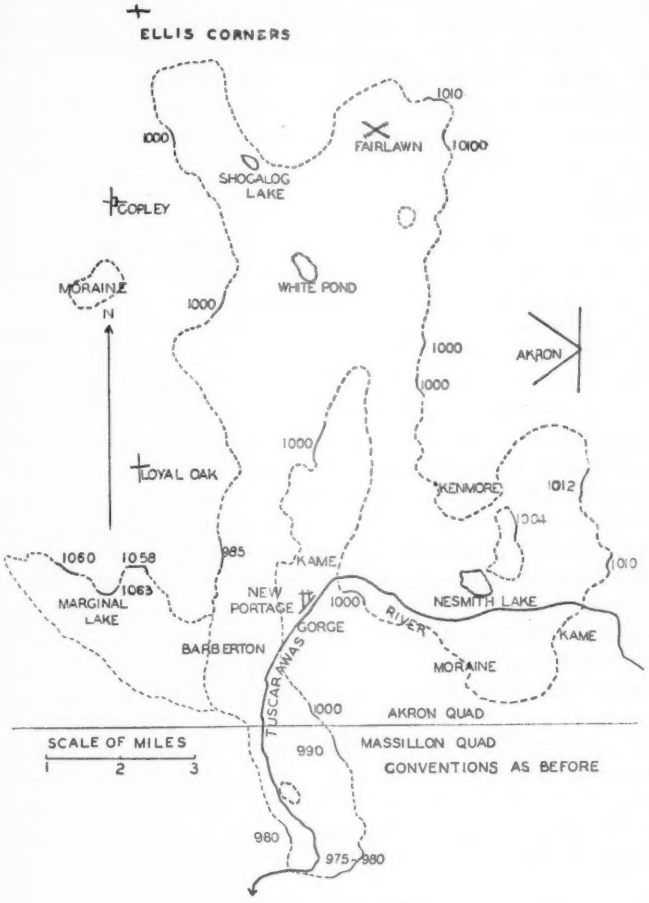


FIGURE 15.—*Akron region*

level, and it continued a mile south and east around the hill. Two groups of kames occur, one 2 miles north and one a mile northwest of Barberton, and no outwash at the surface was found. Outwash in the valley is known to be deeply buried by lake deposits, found in scores of places

over the valley floor. Analyses made but not published show them to be well sorted silts very similar to the histogram (Fig. 14).

Peat is abundant and black earth occurs in many parts of the valley, almost continuously from Barberton to White Pond. More patches occur in the branch valley west of Copley, in the branch reaching northeast toward Fairlawn, in the connecting valley south of Akron and Summit Lake, and in still other small ones south of Barberton. The moraines mentioned cut off two areas, one near Ellis Corners, the other west of Copley, so that they contained small lakes at a slightly higher level than the main lake. In Hudson Run valley an excellent marginal lake, held in by ice, deposited silts to heights of 1060 to 1040 feet. Likewise northwest of Copley a similar lake lay in the valley and left silts up to 1050 or 1055 feet. Peat covers some silt at higher levels here than at any place in the main valley. In two places the main lake was restricted but not dismembered by moraine and kame.

Silts occur beneath the peat and black earth and over all the valley along the margins of the peat. Figure 15 shows the altitudes at which silt occurs and the inclination of 30 to 35 feet in the length of the lake bed.

A mile northwest of Kenmore is a sandstone quarry in a rock-hill slope, which may have been steepened by ice erosion and resteepened by wave work at the 1000-foot level where the silts abut on little sandstone cliffs. Over a mile farther west on a west slope of the next hill are better-marked little sandstone cliffs with silts and sands running up to them on the 1000-foot contour line for more than a mile.

Two broad shallow cols occur at the north end of each north arm of this lake bed, one a mile northwest of Fairlawn which does not now quite reach 1000 feet and one 1½ miles southeast of Ellis Corners which exceeds 1000 feet. Both were carefully studied and yielded no evidence of water standing or flowing over them. A kettle occurs near the west col suggesting no erosion, but it is believed that headward erosion by a tributary stream of Cuyahoga River etching back into the other col has lowered the actual barrier a few feet in postglacial time. The outlet of Akron Lake was through the Tuscarawas River which has cut a gorge into the lake bed at New Portage and which still earlier cut through moraine 4 miles farther south, letting the lake level slowly fall.

### HUDSON-MUD BROOK VALLEY

In the midst of the maze of interlobate moraine between the Grand River and the Black River lobes lies a valley leading southward from the Brandywine, a stream which flows west to the Cuyahoga River near Cuyahoga County south line. Its stream, Mud Brook, rises west of Hudson, flows south through two small lakes (Fig. 16), and gets within

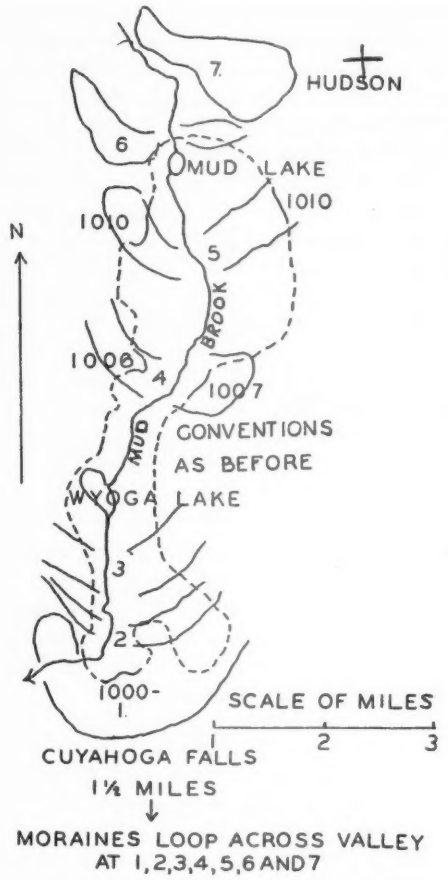


FIGURE 16.—*Hudson-Mud Brook valley*

about 2 miles of Cuyahoga Falls, then turns west, breaks through moraine, and flows steeply down its gorge to the Cuyahoga River. Its turn is favored, probably required, by moraine a trifle higher just to the south of the turn. In this 6 to 7 miles seven feeble moraines loop across the valley, but each except the southern was broken by meltwater erosion when built and not by a through stream.

Silts occur on the moraine that plugs the south end of the valley and southward up to 1040 feet. Silts also are found, isolated from the lake-

bed silts, on the slopes north of Silver and Little Lake as high as 1075 feet, but there is characteristic till at 1080. It looks as if the ice, when building the first or southern moraine in the valley, ponded waters in the region of the above lakes up to the top of the silts, and south of the moraine up to 1040 feet; then, when the ice melted back to let the water out over moraine in the present outlet, the lake level fell to about 1000 feet at the south end. These silts could not have been deposited in water continuous and so high in the valley northward else similar silts would cover the slopes much higher above Wyoga Lake than at present. Short halts in ice retreat were responsible for moraines 2, 3, 4, 5, 6, and 7. No. 7 closed the valley, and there seems to be no evidence of lake in silts and clays north of it. Silts reach 1010 or 1012 feet on the slopes of moraine loops and valley walls south of Mud Lake and 1005 but not 1010 near Turtle or Wyoga Lake. Since these valley silts cover the loop moraines across the valley they represent a lake stage later than the moraine building and the laying of any outwash if present, and their altitudes testify to a 12- to 13-foot slope in 6 miles.

### PYMATUNING CREEK-KINSMAN AREA

This valley was suspected at first of harboring a lake bed perhaps 10 or even 15 miles long. However, silts and clays are limited to 4 miles northward from the moraine at Farmdale P. O. and Kinsman (Fig. 17). The map shows the altitudes of the silt and an inclination southward of about 20 feet in 4 miles. This is not quite in the usual direction of maximum tilt but is the largest rate found. Errors in identifying the top of the silt may make a difference of a foot or two in the total amount of slant.

Several sections were run across the valley both north and south of this 4-mile lake bed, but no silts were found. Drift and fans were common.

The kame of gravel occurs nearly 3 miles north of Kinsman and is being extensively worked now; the silt-laden wash water from the screens spreads over portions of the plain. In this way new yellow clayey silt has been spread 1 to 2 inches thick over many acres. It may perhaps be mistaken for lake silts when the kame is gone and the excavations are healed and forgotten.

### PYMATUNING RESERVOIR

A large marsh called Pymatuning swamp existed along the Ohio-Pennsylvania line from the latitude of Andover 6 miles northward. Recently in the interest of flood and water conservation a large earth dam was thrown across the Shenango River which by a circuitous route

flowed through the swamp. The dam ponds the water, drowns the swamp, and produces a horse-shoe-shaped lake 10 miles long in Ohio on its western arm and 6 or 7 miles long in Pennsylvania on the other. In the summer of 1939 the lake stood at 1008 feet. Although the area was not studied before the water was raised over the marsh, the flooding was not an unmixed evil. No one has ever suspected this valley of having been

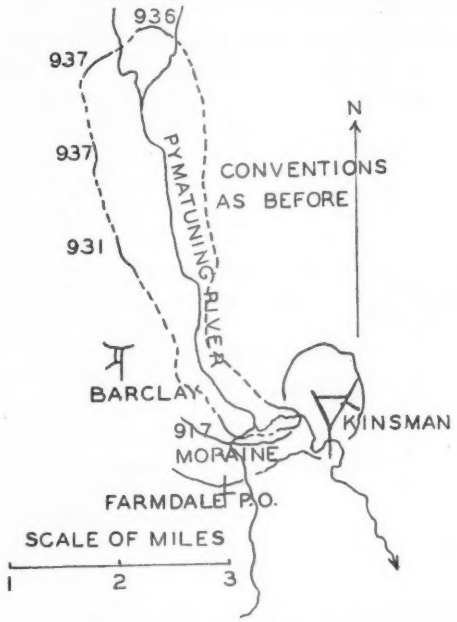


FIGURE 17.—*Pymatuning Creek-Kinsman area*

an outwash channel, and there seem to be no valley dependency glacial features within it. The more definite valleys of Shenango and Crooked Creek, both mostly in Pennsylvania, several miles south of the marsh conditions, do have glacial moraines and other deposits, but these probably have nothing to do with the marsh conditions. Under these circumstances silts could not be confused with moraine terraces, outwash, and glacier milk residues.

Search for silts around the southern end of the west arm was fruitless; but around the northern end, southward along the west side for 4 miles, and in a crescent in Pennsylvania in the angle of the lake silts were found.

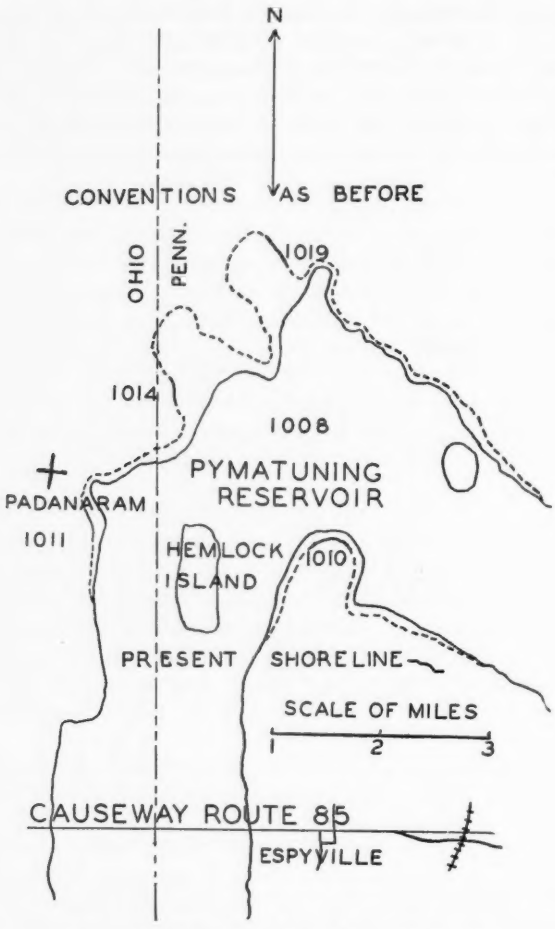


FIGURE 18.—*Pymatuning Reservoir, northern part*

The map (Fig. 18) shows their altitudes in the north up to 11 feet above the present lake level and their gradual descent southward to a point about a mile south of Padanaram where they plunge completely under the lake. The presence of the new lake makes leveling to the edge of the silts a simple and accurate process. The inclination amounts to 11 feet in 4 miles.

Before submergence an island stood in the marsh 2 miles north of the Andover road, now called the causeway, an island called on some maps Hemlock, on others Doré Burt, probably the name of a former owner. This island is now covered with water. Here and toward Linesville were extensive onion and potato farms which farmers described as fine sand, silt, muck, and peaty soils. Probably the Hemlock Island and the farms east were abandoned lake bed lands.

### AREAS INCORRECTLY SUSPECTED OF BEING LAKE BEDS

Several valley floors carefully examined in the hope of finding lake sediments were barren. They are mentioned here for comparison with similar valleys that did contain lakes.

A great advanced mature valley, crossed by route 42 at Five Points about 5½ miles west of Ashland, extends many miles north and south. It was traced south to Perrysville and north along Black Fork of Mohican. This valley is 1½ to 2 miles wide at the top of the filling, at altitude 1000 feet and 4 to 5 miles wide at altitude 1300 feet. It was glaciated by deposition much like Jerome Fork above Ashland, only here the moraine hummocks are not subdued either by lake erosion or deposition, nor are they silt-covered. This absence of silts and clays, and of lake erosion on all moraine hills and valley slopes, is taken as evidence that this valley never had a postglacial lake.

South of Akron and east of Nelson Ledges lie considerable low plain areas several miles in length with small lake beds and occasional lake areas. Drift and alluvium are usually the surface material. The silt and peat areas are small relic kettles.

The Killbuck Creek valley leading from the Lodi marshes south past Wooster, Holmesville, Millersburg, and on south beyond Killbuck has the form and swampy condition of a lake bed in an old valley, but it has no silts and very little peat; nor was any obstruction found sufficient to have held water in the valley. This valley has been twice referred to above and in earlier studies was considered a lake bed valley.

The marshes though abundant are believed to be not remnants of an old lake bed but new features on a flood plain caused by the clearing of hill forests many years ago. This permitted water to hurry down slopes to the valley where no adequate drainage was afforded by clearing the valley or straightening the stream. The histogram (Fig. 19) made from analyses of the sediments taken a few miles south of Millersburg shows how little sorting has been done. This is alluvium, not silt. Several analyses of doubtful silts taken in places where lake beds were suspected have shown as this one does the absence of a lake bed and lake silts.

Wisconsin moraine terraces were found half a mile below Aukerman on the east side at two levels, one about 30, the other about 50 feet above the wagon road. Here and about a mile north, openings have been made in the terrace to remove gravel, but only poorly sorted material is found. These terraces do not continue far, but about 3 miles downstream above a steep rock slope and a terrace of 40 feet is another fragment of terrace consisting of gravelly till.

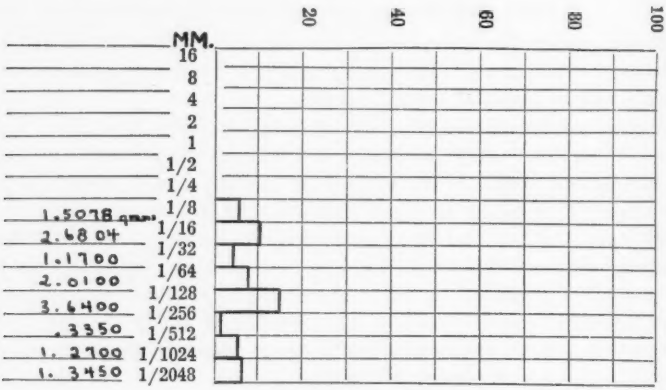


FIGURE 19.—*Histogram of Killbuck Valley alluvium*

Again 4 miles northwest of Wooster on the more gentle slopes south of the east-west road two terraces were found at about 60 and 100 feet above the valley floor. A mile farther toward Wooster just above the two houses, hence about 120 feet above the valley floor, there is a feeble drift terrace, but nowhere was any continuous terrace found, although miles of the valley walls were walked. In places ravine torrents cut through 30 to 40 feet of drift, but usually the main valley walls are only thinly mantled. Gentle slopes were systematically walked for miles in the Wooster region.

In concluding the description of this valley, one can say that moraine terraces are recognized at several isolated places along the valley walls but streams and ice left no continuous terraces. No moraine loops cross the valley below the Holmesville moraine. Although this valley is broad —advanced mature—like the Ashland arm of Craigton Lake valley, it has no such abundance of drift. This valley was glaciated by Wisconsin ice but was given very little drift and not much of its drift is in moraine terraces. Some of the terraces are gravelly and probably stand for local

marginal drainage deposits. The terraces cannot be accepted for shore lines because the valley has no silt over its floor.

Buckeye Lake, made for a canal reservoir 100 years ago, occupies a former swamp area to which early surveyors gave an approximate area of 4000 acres, between 6 and 7 square miles. The present lake and all its fringing marshes do not make so large an area as that. The water is 8 feet higher than the swamp level, and no silts could be found along its borders. It is concluded then that whatever lake there may have been in early postglacial time is completely drowned by the present waters retained by an artificial earth and concrete dam.

There probably are other areas on the Early Wisconsin till sheet that should be investigated in the near future and the results added to this study.

## DISCUSSION

### VALIDITY OF DETERMINATIONS OF SILT

Because challenges usually come on this point when the results have been presented orally a paragraph or two will be added here. In many places the authors have gone along the road or the field in first one direction, then in the opposite, to see if the edge of the silt could be located in any other place. Traverses have been made purposely from obvious silt areas to obvious till areas with no stops, to get the full effect of the change, then with many stops to see the silt thin out or meet the till. Farmers have been asked to characterize their soils on what had been identified as silt and till. They have been asked where the boundary is between the silt soil and the till soil. The soils cultivate so differently that most farmers put the boundary about as it has been mapped. Where gradations occur it is assumed that the purer the silt, the less clay it contains, and that it may grade into very clayey material toward the central part of the lake or downward in the central part, or that it may grade into more clayey material as one goes from lake bed to till sheet. In the one case, silt becomes rarer with increase of clay but with few or no pebbles. In the other case, the silt decreases with increase of clay and of sand, pebbles, and boulders. In the first case the sorting continues excellent with grading from silt to clay, while in the latter the sorting fails rather abruptly when the silt-till boundary is passed. In very few places has the margin of the silt been marked by gravel or sand that could possibly be outwash or moraine terraces. Beaches are rare. The shapes of most of the lakes are not consistent with interpreting their deposits as outwash. Also lake positions in most cases are inconsistent with such interpretation. Many of the lake beds are on till sheets and not in elongated valleys. Those in elongated north-south valleys having the same symptoms and characteristics are included in this paper.

The clayeyness is estimated in part by the irridescence of a smooth surface. Silt has almost no play of colors; "jackwax" has splendid displays. In till a sheen is present, but sand and pebbles scratch a knife, interfere with whittling the clod, and show with frequency under a hand lens magnifying 10 to 14 times. The clay beds, clay-silt, and much of the silt contain brown to black concretionary grains like tiny flax seeds. These are richer in iron than the matrix material; they are not in layers, are easily cut with a knife, are not crystalline, and occasionally show definite concentric structure. These concretionary bodies are not found in till, hence they are deemed characteristic of silt and not related to a soil profile.

In some fine silts and clays the varves were clear and definite enough so that they could be pulled out on an auger. If not too deeply covered, varves were dug out with the spade and trimmed to make them show on a smooth side in the hole. Varves may not always be seasonal but are caused sometimes by winds. They indicate lakes. Presumably most of the lakes had varves when the beds were made, but roots of forest trees and deep cultivation have damaged them in the surface layers. No doubt worms and other burrowers have destroyed them in places. Therefore it was not disturbing if clays and silts did not show much bedding in the 10 feet that the auger would reach.

For field-work purposes silt has been defined as that material whose grains are not visible to the naked eye but can be seen with a magnification of 10 to 14 times. This ruled out the sands on one side and the clays on the other. The material was often handled both wet and dry to test its coherence and its balling and pulverizing qualities. Likewise effort has always been made to keep in mind distinctions between alluvium and lake sediments, between outwash and all other meltwater deposits and those made as beaches or bars near lake borders, and between more or less sorted drift and true lake silts. Zones in weathered soil profiles and silts have often been compared. Wind-blown material, while constantly watched for, has rarely been found, probably because lakes were small and usually filled with vegetation before they became dry. Thus they lost their water very gradually, leaving dead and growing vegetation over the silts. They were not emptied quickly by the discovery of a new lower outlet as were the Maumee and Whittlesey beds in the Great Lakes area, but were lowered by the cutting down of the outlet.

### RELATION OF LAKE BEDS TO MORAINES

The lake beds herein described are closely associated with the Late Wisconsin moraine system. They are largely between the first, Union,

and second, Mississinawa, moraines; but Indian Lake is just south of the first, and St. Henry, Marseilles, Scioto, Willard, Hog Creek, and Hudson are north of the second. This distribution, together with the youth of the lake beds, is eloquent testimony to the recency of Late Wisconsin as well as to a considerable interval between the last of Early and the first of Late Wisconsin depositional work. If there were similar lakes and then abandoned lake beds beyond the influence of Late Wisconsin, the evidence of them has been nearly obliterated by oxidation of their peat and dissection of their plains. Dachnowski (1912, p. 49-51) cites two in Darke County and comments on the disintegrated condition of the peat and its possible greater former extent. He also mentions small patches of peat in the interlobate moraine of Clark and Champaign counties, and in several other places.

It then remains that the lake beds which are relatively recent and large enough to contribute to this study are distributed closely along the Late Wisconsin border. Their genesis is much later than any that may have existed on Early Wisconsin or Illinoian drift areas, and all but the Willard are distinctly earlier than the beginnings of the Maumee Lake. All are considerably higher than the Maumee beach and nearer the State divide; some lie on the divide, some are south of it but quite near.

### THE PROBLEM OF TILTING

Several of these lake beds, as previously stated, were examined by the senior author and reported before the inclination of the deposits and shore features was discovered. With detailed mapping of Craigton Lake it became very obvious that shore features were not level as they must have been when made. The mapping showed that there is a rather uniform slope to them. They do not step down as if they represented repeated withdrawal of ice or shiftings of outlet. This came out even more clearly when the mapping was done on the silts.

The study has now been extended to include all lake beds that could be found in the region south of the Great Lakes area. The exploration has been carried much farther south than lake beds could be found to be sure that nothing was omitted which should be included. If omissions occur they are still farther south on older drifts.

It seems safe to assume that silt and clay in a lake bed would be deposited up to essentially uniform altitudes over all the bed and that they could not be deposited higher than the lake level. The upper limits of the silt have been found to be, in these 13 lake beds, higher on the north and northeast than on the south and southwest margins. When this condition was discovered the authors were driven to the conclusion

that the lake beds had been tilted. Were one working with actual shore lines instead of silt, clay, and peat sheets across lake beds, he could say shore lines are tilted, but shore lines are not the best evidence of tilting here.

The tilt is present all across the State from west to east. It is possibly a little greater toward the east, for if the tilts of five eastern beds be averaged and the rest be averaged separately, the eastern tilt is about 3½ feet per mile while the western is a little less than 2 feet per mile.

Further, the tilt as found is as gentle as 1 foot, once 5 inches per mile, and in one case reversed; or as steep as 2 or 3 feet per mile, and in some cases as high as 4, 4½, even 5 feet. The apparent inclination of the silts in the reverse direction was found in but one lake bed in 14 and that is the small one about 3 miles across. It has been drained and farmed a long time, and its silts never were thick.

Table 1 shows in each of the 13 lake beds the amount of tilt, the distance across, and rate of tilt per mile.

TABLE 1.—*Tilt in 13 lake beds*

| Lake | Tilt (Feet) | Distance (Miles) | Rate per mile (Feet) |
|---|---|---|---|
| St. Henry | 5 | 3 | 1⅔ |
| Indian | 10 | 5 | 2 |
| Holden | 4 | 3½ | 1 |
| Scioto | 3 | 7 | .5 |
| Hog Creek | 3 | 3 | 1 |
| Marseilles | 17 | 4 | 4 |
| Marion | 10 | 6 | 1⅔ |
| Shelby | 4 | 3 | 1.3 |
| Craig | 80 | 18 | 4.5 |
| Akron | 20 | 10 | 2 |
| Hudson | 10 | 5 | 2 |
| Pymatuning Creek | 20 | 4 | 5 |
| Pymatuning Reservoir | 11 | 3½ | 3.5 |
| Totals | 197 | 75 | Ave. 2.6+ |

The fact of tilt is much more certain than the actual amount in any one case, and the direction of tilt, up in the north, is more certain than the uniformity of tilt. This belt of former lakes is 30 to 35 miles wide north and south. The average tilt is over 2½ feet. The total tilt may be 75 to 85 feet across the belt.

### CORRELATION WITH TILTING IN GREAT LAKES BASIN

Correlations with actual shoreline tilts in the Great Lakes area are suggested by the following: (1) The general direction is much the same, (2) the average amount is much the same, (3) the greater tilts per mile are toward the east, (4) it seems reasonable to suppose that the cause of

tilt is the same as that invoked in the Great Lakes area—the isostatic adjustment and the resiliency of the materials of the earth and their unloading by melting of the continental glacier.

## SUMMARY AND CONCLUSIONS

It is believed that the uplift was not uniform, not even uniformly increasing northward, and in some places was more easterly than in others; but it is not believed that the uplift was as uneven as the actual figures suggest. Probably the zero line, or zone, in tilting was near the ice margin for each time involved: not at the Illinoian border for these Late Wisconsin beds, but more nearly where the ice border was in the beginning of Late Wisconsin retreat. All hinge lines ever mapped have been placed near the edge of the ice and water removed before the tilting occurred. A hinge line was hypothetically placed by Fairchild (1918a; 1918b) and by Baron De Geer (1892; 1893) far south of any of these lake beds.

There may well have been tilt in southwestern Ohio following the withdrawal of Illinoian ice and again when Early Wisconsin ice melted, and perhaps these tilts followed back successively somewhat like a wave as the several ice loads melted off.

It seems probable that the tilting and uplift were approximately commensurate with depression caused by advance of the ice. It is strongly suspected that depression and resilient uplift and tilting are related not only to load, ice, and water, but to strength of materials in the substructures of the earth.

These lake beds were probably tilted after their drainage and filling had begun, as were the abandoned shore lines of the Great Lakes area. Probably the tilting in central Ohio was well under way before that in the Great Lakes area began. This would be in harmony with such recovery as has been noted, but better evidence is in the fact that successive hinge lines have been found in the Great Lakes area, where old beaches can be followed many miles.

No evidence has been found as to whether these lake beds are still being tilted or not, but, by analogy in the Great Lakes area, it is strongly suspected that the tilting is past, even though it may still be in feeble progress in the Great Lakes area.

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Oberlin College, Oberlin, Ohio

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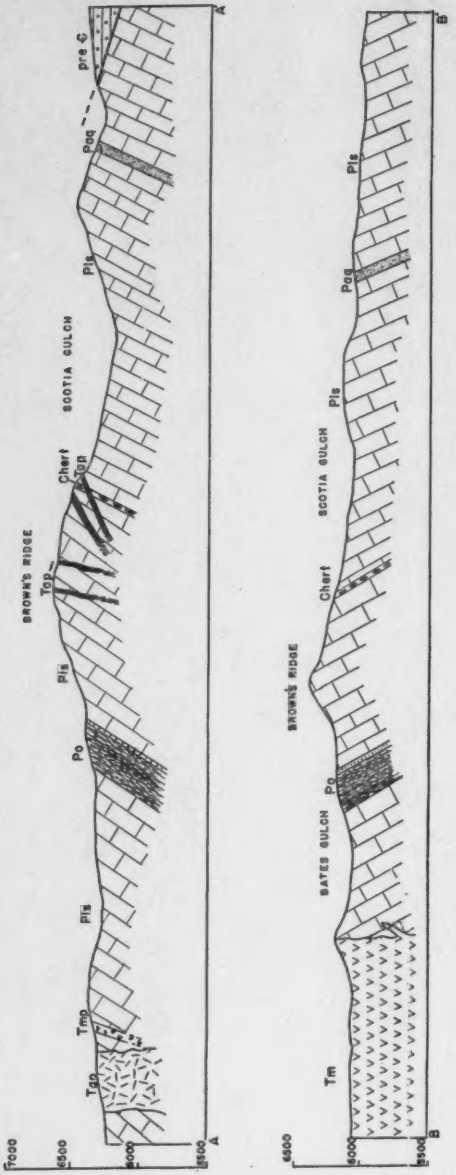


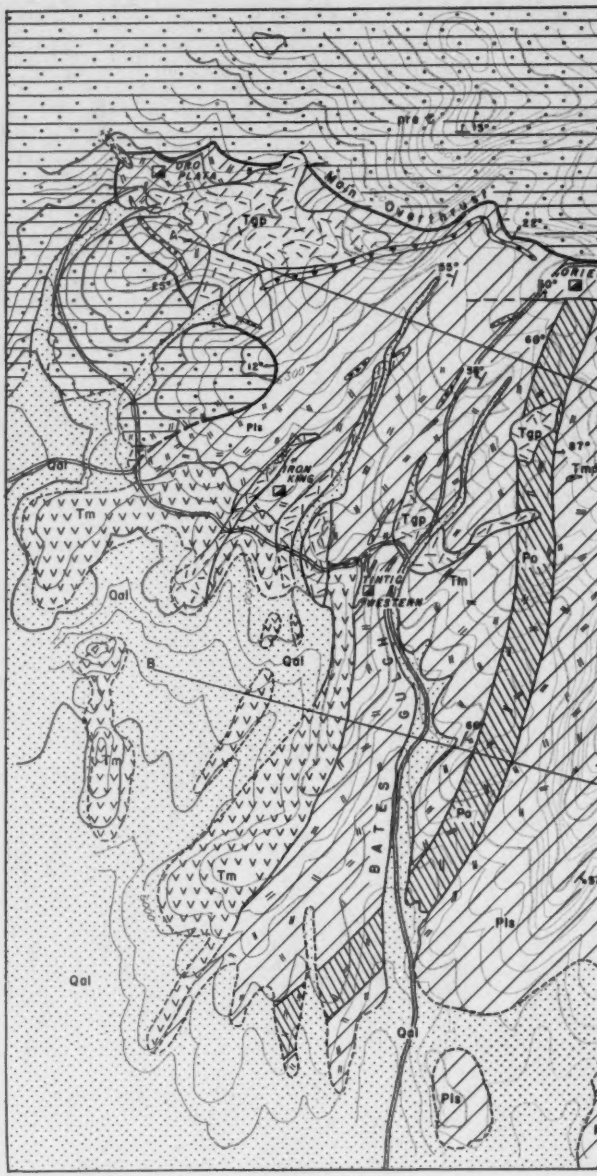


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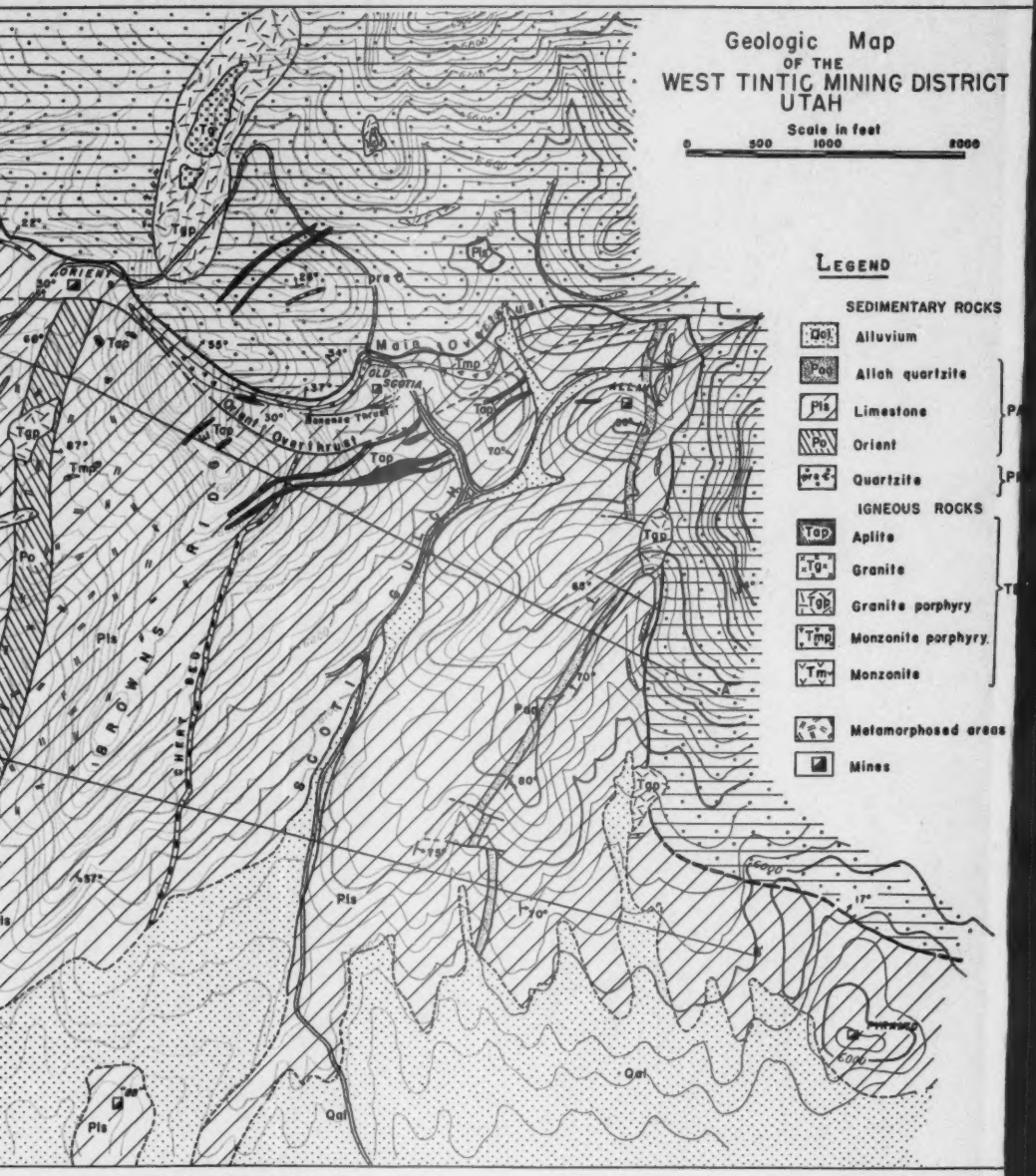
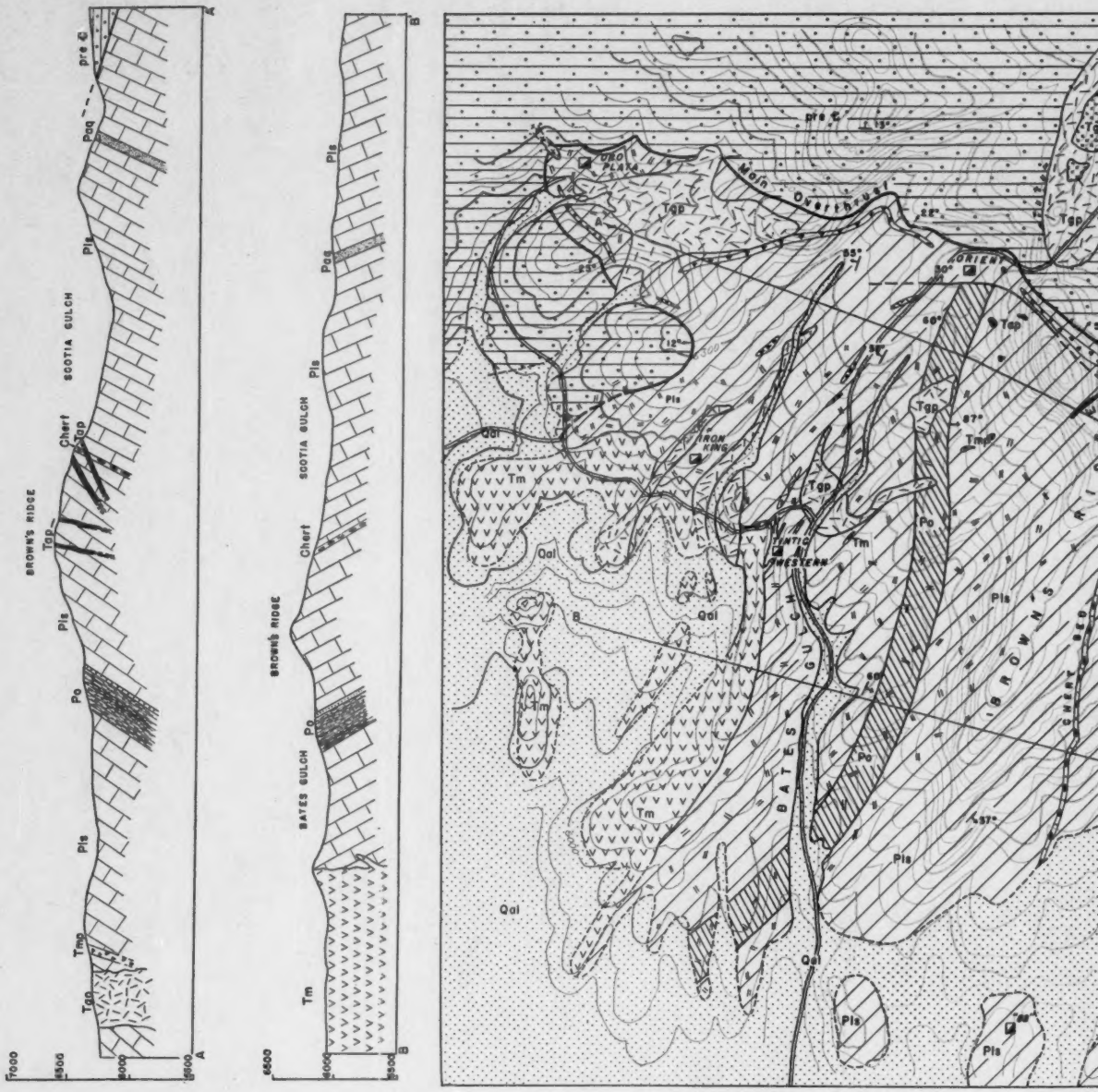
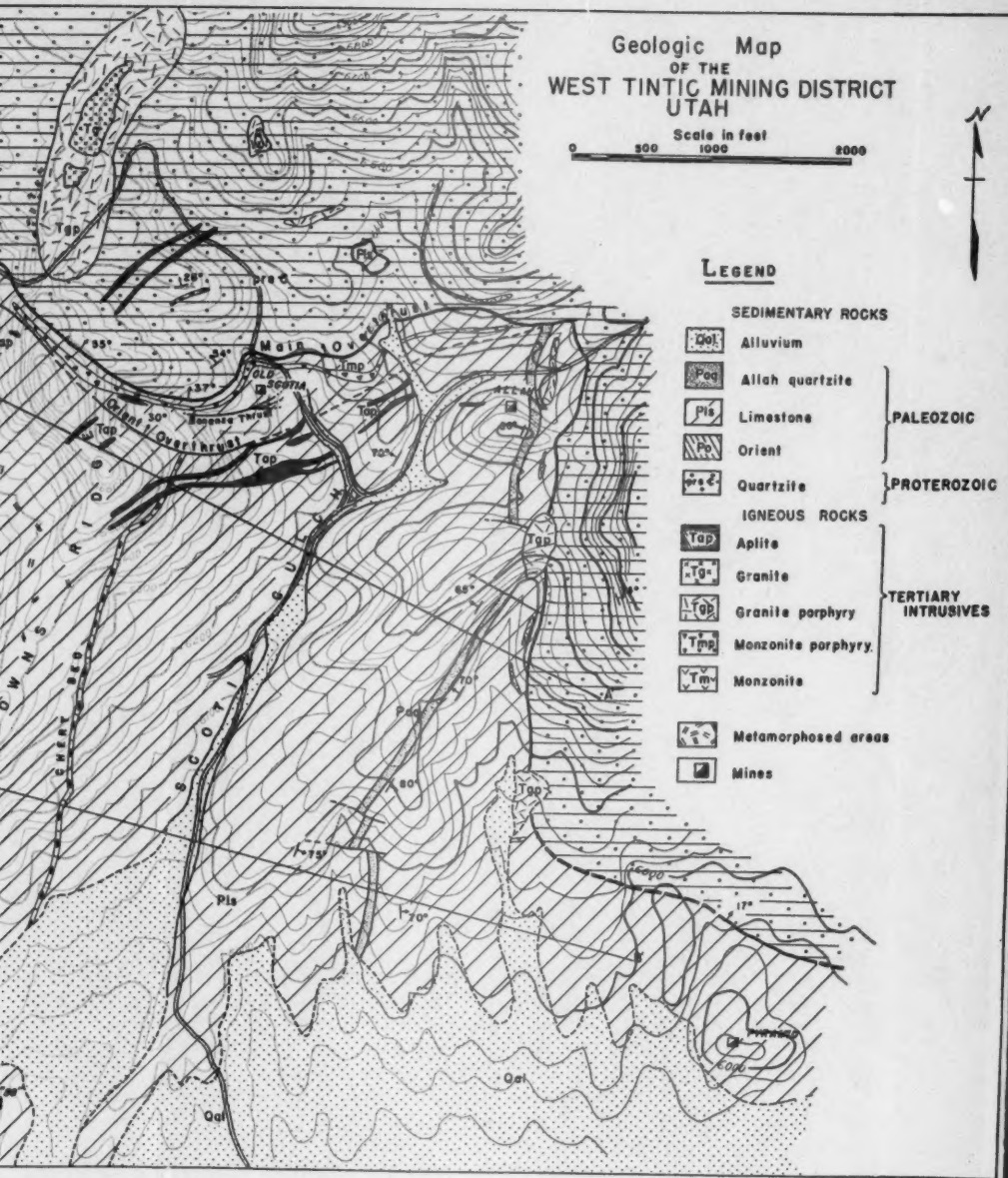



THE WEST TINTIC MINING DISTRICT, UTAH




GEOLOGY AND TOPOGRAPHY OF THE WEST TINTIC

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# MINERALIZATION IN THE WEST TINTIC MINING DISTRICT, UTAH

BY BRONSON F. STRINGHAM

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## ABSTRACT

In the West Tintic Mining District Proterozoic quartzites and slates have been thrust over folded Paleozoic limestones and dolomites. Five types of igneous rocks have been intruded in sequence of four stages: (1) monzonite stock simultaneous with monzonite porphyry dikes, (2) granite porphyry stocks with (3) granite cores, (4) aplite dikes.

Contact metamorphism is prominent adjacent to the larger intrusions and is considered to have been accomplished by end-stage liquids with little addition. Pyrometasomatic veins and mesothermal deposits were next formed and were followed by the formation of low-temperature quartz-barite veins and replacements. The entire sequence of events is established by field and laboratory evidence.

It is believed that here is represented the history of a cooling differentiating magma and the deposition of hydrothermal minerals of progressively lower temperature of formation.


## INTRODUCTION AND ACKNOWLEDGMENTS

The West Tintic Mining District is approximately 21 miles southwest of Eureka, Utah. The original area included in the district is somewhat indefinite. The principal mines and prospects, however, are confined to about 4 square miles of limestone area at the southern margin of the Sheeprock Range. It is best approached by a road from the south which branches from Utah State Highway No. 26 at Jericho and proceeds westward around the southern end of the West Tintic Range across Cherry Creek and thence to the district (Fig. 1).

First shipments were made in 1870 when a small pocket of high-grade gold, silver, and lead ore was discovered at the surface near what is now known as the Old Scotia Mine. This discovery stimulated prospecting for several years, and many prospect pits and shallow shafts were sunk. Soon after the discovery, the Old Scotia developed another small ore body which has since been worked intermittently. There is no information available concerning the dates of other notable developments. A new prospect, the Tintic Western, was operating in the summer of 1940.

The district has nearly escaped notice by geologists, probably because of its limited production. Only one published account is available (Loughlin, 1920). The report includes a reconnaissance map and a brief description of the geology and ore deposits. The present study confirms much of Loughlin's work but adds considerably to its refinement.

During the summers of 1936 and 1940, the author devoted a total of 4 months to field study. Detailed topographic and geologic maps were made, and two mines were mapped. This has been supplemented by laboratory work.

The writer wishes to express his appreciation to Professor Paul F. Kerr of Columbia University for his encouragement and advice during the study of this problem, both in the field and laboratory. Thanks

are also extended to Professors S. J. Shand and William M. Agar, also of Columbia, who read the manuscript and offered valuable criticism.

## TOPOGRAPHY

The Sheeprock Range is a maturely dissected block mountain typical of those in the Great Basin. The West Tintic District is situated at its southern end at comparatively low elevation. To the north at higher elevations, steep slopes, a few small cliffs, and rugged outcrops may be seen. Toward the southern part, however, are low, smooth, rolling hills, mainly covered with alluvium. A prominent feature is a high symmetrical ridge extending entirely across the district from north to south. Its summit slopes gently southward but steepens abruptly near the southern border of the mapped area. Paralleling the ridge on either side are valleys which have been partially filled with alluvium. For convenience the ridge has been named Brown's Ridge, the valley to the east Scotia Gulch, and the one to the west Bates Gulch.

## SEDIMENTARY ROCKS

### PROTEROZOIC

As the result of overthrusting Proterozoic rocks lie above Paleozoic. The former consist of interbedded quartzite, slate, phyllite, and conglomeratic slate. Recrystallized quartz grains of fairly uniform size compose the bulk of the quartzites with a little chlorite and sericite as impurities. Many grains are broken and brecciated indicating incipient cataclastic metamorphism. The argillaceous rocks are slaty, but a few beds contain mica and chlorite so well developed that these should be more properly classed as phyllites. They are usually greenish but grade into purple and black. Several slate units contain subrounded boulders of quartzite, granite, schist, and older slate and have been likened to tillite by Loughlin (1920, p. 432). Thicknesses of lithologic units range from a few feet to several hundred feet. No particular sequence of beds can be observed. Almost every outcrop of quartzite in the Proterozoic area shows extensive slickensiding. Moderate dips up to 34° are present, varying in direction from west of north to northeast. The general structure is complicated by many faults of small displacement. These do not extend into underlying Paleozoic formations and hence probably antedate the thrusting. Many pure, white, massive quartz veins traverse the formation, in some places producing a white network. Rare limonite stains are the only evidence of any other mineral accompanying the quartz. These veins evidently antedate the thrusting

because they are also slickensided and exhibit few of the characteristics of the quartz veins in the limestone area.

Little time was spent in attempting to unravel details of the Protero-

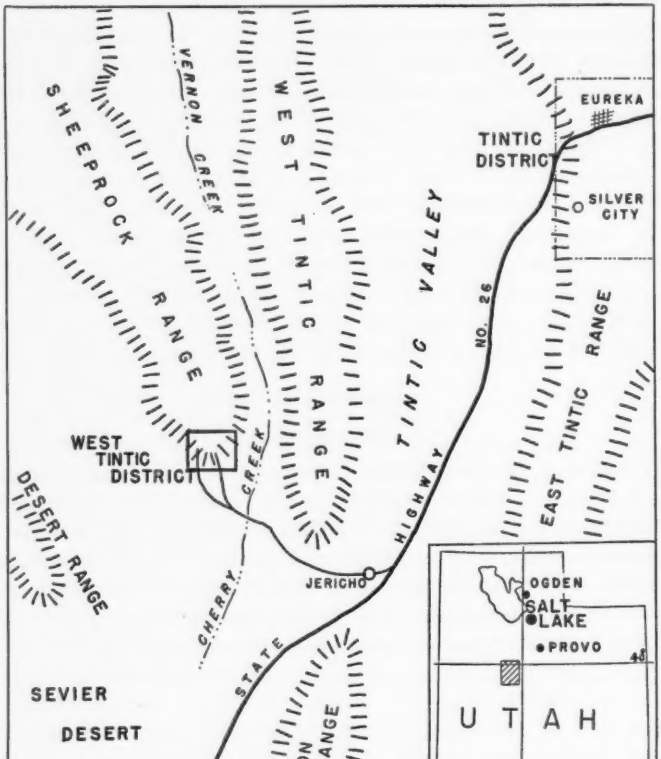


FIGURE 1.—*Locality map*

zoic rocks, since their structure and mineralization appear to be entirely unrelated to the Paleozoic sediments.

Loughlin (1920, p. 432) questions the age of these rocks by saying:

> "There is no local evidence to indicate the age of this formation, but as it is continuous northward . . . where it underlies quartzite of Lower and Middle Cambrian age there can be no reasonable doubt that it is either very early Lower Cambrian or pre-Cambrian."

Eardley and Hatch (1940, p. 826) have described and measured sections of this same sequence to the north, and there is little doubt that these rocks fit well down in their column of Proterozoic age far below the Cambrian.

### PALEOZOIC

An estimated thickness of 7000 feet of limestone and dolomite makes up the greater part of the sedimentary rocks exposed below the overthrust. With the exception of four horizons, distinct lithologic units are not sharply defined. An attempt is made, however, briefly to describe broad units and give their positions without regard to specific thicknesses.

In the southeastern part of the area massive blue coarse crystalline dolomites predominate in which bedding is obscure, but occasional chert lenses give a clue to the general structure. West of this is a 50-foot bed of fine-grained quartzite with prominent bedding planes. Because of its proximity to the Allah mine, it is here called the Allah quartzite. This bed is easily traceable from north to south across the entire district. West of this and continuing nearly to the Scotia Gulch is a series of alternating gray and blue massive limestones and dolomites containing a little sand. On either side of Scotia Gulch is found a well-stratified cream-colored sublithographic limestone. West of this and exposed in Brown's Ridge is a series of alternating blue and gray coarsely crystalline dolomitic limestones with a traceable bed of chert about 20 feet thick. Several fossils were found on the west side of the ridge from which *Actinoceras* sp., *Halysites catenularia*, and *Streptelasma* sp. were identified as Ordovician by Dr. H. N. Coryell of Columbia University. In contact with these rocks to the west is 15 feet of brown, well-crystallized quartzite and a 250-foot unit of dense, greenish-brown shale. Both of these are easily traced from the Orient Mine south and westward across Bates Gulch. These will be referred to collectively as the Orient formation and specifically as the Orient quartzite and Orient shale. Very little unmetamorphosed sedimentary rock is present west of this horizon, but a few fresh outcrops are predominantly buff limestone with thin lenses of argillaceous and arenaceous material which, when weathered, impart a laminated appearance.

An argillaceous shaly limestone of limited extent occurs near the Old Scotia Mine. Its 60-degree dip to the north indicates that it is entirely unrelated structurally to the rest of the Paleozoic rocks. This formation contained the bonanza ore of the early days and was named the Bonanza shale.

### QUATERNARY

The Quaternary alluvium is composed of unsorted silt, sand, and rock fragments of all shapes and sizes up to a foot or more in diameter. In the northern part of the area it covers only the bottom of valleys having low gradients. Here the alluvial boundaries are distinct enough to map. To the south and west at lower elevations, however, all slopes are covered, including the crests of low ridges, and consequently it is difficult to map the alluvium with a definite boundary. Information gathered from observations on the few available bedrock outcrops is therefore insufficient for good interpretation. Hence the study of the mineralization has been limited as far as possible to the northern areas where good exposures furnish superior evidence.

## STRUCTURE

### STRUCTURE OF PALEOZOIC ROCKS

The Paleozoic strata form a steeply dipping homocline which trends north and south. This conclusion is arrived at from the fact that the Allah quartzite, the Orient formation, and the intervening chert bed can be traced from the northern part of the area south to the alluvium, and also that there is no repetition of lithologic units anywhere in the district. The variable dips, however, show that there are complications to this simple structure. In the northern part, dips are to the west. In the central portions the strata stand vertical or nearly so, while in the south dips are eastward (Pl. 1). Overturning is evident, therefore, but to make sure of this the Allah quartzite was traced and dips checked throughout its entire length. It has not been determined which part has been overturned since fossils, joints, lithologic sequence, primary structures, and other features furnish insufficient evidence to determine top or bottom of the beds. On the eastern nose of the central section of Brown's Ridge, a small local flexure involving no great thickness of strata slightly disturbs the regularity of the strike of the formations. (*See* chert bed, Plate 1.) Several minor dip faults offset the Allah quartzite but could not be traced any great distance because of the massive character of adjacent strata. Other minor dip faults involve the Orient formation and chert bed. Mineralization except by calcite which cements breccia is essentially absent along these fractures.

### OVERTHRUSTS

Loughlin (1920, p. 435) gives conclusive proof that the Proterozoic has been overthrust upon Paleozoic rocks. The observable fault zone appears to vary in width depending on the kind of rock overlying. An adit examined northeast of Brown's Ridge penetrated the fault where

FIGURE 1. LOOKING WEST TOWARD THE OLD SCOTIA MINE
Shows the position of the main overthrust.

FIGURE 2. MARMORITE
Laminae of idocrase (dark streaks) and wollastonite (light streaks) parallel to the bedding. Lower part composed of coarse calcite.

FIGURE 3. ANDRADITE VEIN IN WHITE MARBLE
Dark spots in marble are introduced andradite crystals.

PHOTOGRAPHS OF OVERTHRUST, MARMORITE AND GARNET VEIN

quartzite is in contact with the limestone, with 18 feet of loose limestone and quartzite breccia between. Near the Old Scotia Mine slate is in contact with limestone with a 2-foot zone of loose breccia. At a very few places only, along the entire trace of the main overthrust, is there any evidence of mineralization. At some places accurate measurement of the dip of the fault is possible. Such opportunities as were afforded showed the thrust plane to be markedly irregular, especially if it were once to pass over Brown's Ridge where only limestone is exposed (Pl. 2, fig. 1). Erosion has cut through the quartzite cover and exposed a small area of limestone northwest of the Allah mine. Here the fault dips gently southward. It is believed that the movement of the upper block was over a very irregular surface since the Paleozoic strata show no evidence of having been deformed after thrusting. The extreme development of slickensides in the quartzite may be accounted for by its constant adjustment to this surface during movement.

Other smaller thrust faults cut the rocks below the main overthrust. One is exposed in the Old Glory Hole of the original discovery. It dips 29° N. and can be traced clearly in the upper workings of the Old Scotia Mine. Here the Bonanza shale, which dips 60° N., overlies a massive limestone. This fault can be traced westward only a short distance; toward the east it appears to branch into several small breaks where the structure is extremely complicated. Just south of the Orient Mine, another thrust restricts the northward extension of the Orient formation. This can be fairly well traced on the surface eastward to near the Old Scotia Mine where it cuts the chert bed. It is proposed to name the former fault the Bonanza thrust and the latter the Orient thrust (Fig. 5). Apparently there has been considerable movement along these two faults since beds on either side cannot be matched. In the Old Scotia, however, several gently dipping breaks with small throw have been found. It can be seen (Pl. 1) that the dikes in the limestone between the Old Scotia and Allah mines follow much the same outcrop pattern as the main overthrust. Field and underground examination shows that many of these dikes nearly parallel the thrust in vertical section also. It is supposed, therefore, that they were intruded along breaks in the limestone, produced by the overthrust, some of which have little or no throw. Further work underground above the Orient overthrust may be expected to encounter more breaks and offsets of low dip.

In the Old Scotia mine are large north-south vertical open fissures whose walls are thickly coated with travertine. Ground water has so widened several of these that they are of cave proportions.

## IGNEOUS ROCKS

### MONZONITE

One body of monzonite in the southwestern portion of the district was recognized and described by Loughlin (1920, p. 436) who called it a "stock-like body of monzonite" and assigned it to the Tertiary. The northern and eastern boundaries were traced fairly well by float and scattered outcrops, but to the southwest alluvium effectively obscures the bedrock. The southwestward extent of the stock, therefore, is not known, but outcrops on a few small hills to the west, which are isolated by alluvium, indicate that it is quite large. It may be a mile or two in diameter or may extend many miles out underneath the alluvium of Sevier desert (Fig. 1).

The typical monzonite has a uniform medium-grained granitoid texture. White plagioclase, light-pink orthoclase, biotite, and small crystals of hornblende are discernible in hand specimen. In all of several specimens studied microscopically it was found that orthoclase predominates slightly over plagioclase ($An_{32}$), quartz varies from 8 to 17 per cent of the whole rock, and hornblende remains constant at around 5 to 6 per cent. Medium-sized plates of biotite constitute 10 to 12 per cent. Sphene, augite, magnetite, and apatite are the accessories. Chlorite, magnetite, sericite, and epidote, present in all specimens in minor amounts, indicate that alteration has been slight. The typical texture is seriate with the largest crystals 3 to 4 mm. in greatest dimension. Most of these are prominently zoned plagioclase with twinning on the Carlsbad, Albite, Manebach, and Pericline laws. Determination of the compositions of the zones showed that the anorthite content decreases toward the rim of the crystal. Some cores contained as high as 67 per cent anorthite, and rims as low as 10 per cent. A few crystals of orthoclase are also present. The boundaries of these larger crystals tend toward idiomorphism. Medium-sized hypidiomorphic crystals of perthitic orthoclase, plagioclase, and quartz constitute the bulk of the rock. The crystals of hornblende may show either good crystal outlines or ragged edges. Occasional remnants of augite within hornblende crystals suggest that it has a uralitic origin.

Immediately south of the Tintic Western Mine is a small area of dioritic rock. This differs from the monzonite in having much more hornblende and slightly less orthoclase. Its connection with the main body was not established, but it is believed to be a small apophysis or more basic phase of the monzonite.

### MONZONITE PORPHYRY

The monzonite porphyry occurs principally as dikes and sills not over 10 to 20 feet thick, except for local thickening where outcrops 50 to 60 feet across have been found. Several dikes of limited extent which trend northeast were mapped near the monzonite stocks.

The monzonite porphyry is rather striking. White feldspar and black mica phenocrysts which constitute between 40 and 50 per cent of the

rock are set in a dense, greenish groundmass. The largest phenocrysts are white feldspar, 5 to 6 mm. in greatest dimension. Biotite phenocrysts are much smaller, however, and vary considerably in size.

Under the microscope the larger phenocrysts are found to be well-twinned, idiomorphic plagioclase ($An_{39}$) with little or no zoning, while many smaller crystals have ragged edges. Large crystals of dark-brown biotite have been altered to chlorite around their edges, and smaller ones are wholly chloritized. The groundmass is composed of interlocking crystals of orthoclase, chlorite after biotite, and some quartz. Various stages in alteration are found in specimens taken from different places. Sericite, chlorite, and carbonate are always present. Zoisite and epidote appear only occasionally. An exposure in the excavation for a powder house near the Old Scotia Mine shows the porphyry to have been fractured and sericite and carbonate developed to such an extent that only remnants of the green porphyry are left. Hydrothermal action explains most of these alterations.

The age relationship of the porphyry to the monzonite is not evident in the field since no contacts were observed, but the similar nature of these two rocks strongly suggests that they are of contemporaneous origin. The porphyry bodies probably represent offshoots from the main magma chamber before its consolidation.

Loughlin (1920, p. 434) described a pink granite which he found as an inclusion in a monzonite porphyry dike. This outcrop was not found by the writer, but many boulders of pink granite fitting his description were observed lying always within areas of monzonite porphyry. Loughlin (1920, p. 434) considered them broken fragments of a pre-Cambrian granite, floated up in the monzonite porphyry magma.

### GRANITE PORPHYRY

Granite porphry outcrops in oval, elongate, or irregular areas of different sizes. The largest of the latter is a mass near the Oro Plata Mine and is here called the Oro Plata granite porphyry. Several elongated and irregular masses are grouped near and in the monzonite stock. Northeast of the Orient Mine (Pl. 1) is a large ovate outcrop cored by granite and will be called the Orient granite porphyry. A very much smaller body with a granite core occurs on a ridge directly north of the Old Scotia Mine. Many smaller bodies may be found along the eastern margin of the limestone area very close to the overthrust contact.

In hand specimen this rock is dense, buff-colored, and massive. The true porphyritic character is not immediately evident on casual observation, but phenocrysts of quartz, biotite, and feldspar can be seen when it is examined closely. The phenocrysts vary in amount but never predominate over the groundmass. In a few cases, flowage lines can be seen, particularly in the Orient granite porphyry. Many weathered

specimens are stained brown from iron oxide, but this is only surficial. On fresh breaks many brown cubic masses are present, indicating the former presence of pyrite.

Although all the rocks mapped as granite porphyry have a similar appearance in hand specimen, two marked textural varieties are distinguishable under the microscope. The one nearest to that usually ascribed to typical granite porphyry (Johannsen, 1932, p. 294) is more common in the western intrusive. It is markedly porphyritic with plagioclase ($An_{35}$) composing five-sixths of the phenocrysts and biotite and orthoclase the remainder, all of which total 20 per cent of the rock. The plagioclase is clear, broadly twinned, and well shaped, with dimensions up to 2 mm. Many crystals have rims of orthoclase. Orthoclase phenocrysts are not so large but are clear and twinned according to the Carlsbad law. Biotite is occasionally quite fresh but usually slightly altered to chlorite. The groundmass is composed of intimately interlocking quartz and orthoclase and occasional graphic intergrowths. Accessories are sphene and limonite after pyrite. A few small plagioclase crystals ($An_0$) having about the same dimensions and interlocking relations as the quartz and orthoclase suggest very strongly that the plagioclase phenocrysts formed very early.

The other textural variety occurs in the Orient porphyry and predominantly in the bodies of the eastern part of the district. Thirty to forty per cent of this rock is composed of quartz, orthoclase, and plagioclase ($An_{28}$) phenocrysts. Many crystals show ragged outlines as though they had been formed early and were fractured and broken during flowage before emplacement. The groundmass can be resolved only under high power. It consists of small crystals of quartz and orthoclase with many shreds of bleached biotite. Sericite is prominent in all the rocks of this type. In some specimens pyrite has been introduced profusely.

### GRANITE

The largest body of granite lies entirely within the Orient granite porphyry forming a core of that mass. It outcrops in two areas, separated by about 60 feet of porphyry. The contact between the two rocks is sharp, but the granite cuts the flow lines of the porphyry, indicating that the granite is later. The rock is massive and coarsely crystalline with crystals up to 4 or 5 mm. across. It is normally cream-colored, but much limonite locally stains the rock brown, especially near small fractures where it appears that pyrite has been abundant. The rock is so altered that even fresh specimens tend to crumble in the hand.

The texture is granitoid. Plagioclase ($An_{15}$) constitutes the smaller crystals and has been extensively altered to sericite. Orthoclase is prominently perthitic and turbid with kaolin, while quartz is clear. Biotite and pyrite are the only other minerals and together make up only 1 per cent of the whole. At the northwest corner of the limestone area near the Oro Plata Mine is a northwest-trending dike of muscovite granite. Loughlin (1920, p. 435) described it as a "nearly white, fine, even-grained rock composed of white feldspar, colorless quartz, thinly scattered muscovite and some flakes of biotite." It is less altered than the other granite and differs only in having a little muscovite, slightly more basic plagioclase, practically no pyrite, and a finer texture.

### APLITE

The thin dikes, sills, and small irregular pipes of aplite are confined to areas close to the overthrust fault between the Orient and Allah mines. Areas near the Orient Mine are small and roughly circular. Southeast of the Old Scotia the outcrop roughly parallels the overthrust contact. In this mine the dikes dip gently northward, which is also parallel to the dip of the fault. The thickest dike measured underground is 12 feet; consequently on the map the width of most of them is exaggerated. However, a few large exposures offer proof that they thicken locally. One small dike, 6 inches wide, was found cutting the Orient granite.

In hand specimen the rock is hardly distinguishable from the granite porphyry except for the phenocrysts. It is buff to tan-colored and fine-grained and locally stained light brown by limonite. Fractures and flow lines are usually prominent and are parallel to the walls of the intrusion. West of the Old Scotia the aplite is found adjacent to monzonite porphyry, but the contact exhibited no criteria for determining the sequence of intrusion.

Under the microscope, it is seen that quartz and orthoclase make up almost the entire rock. Biotite in its greatest concentration constitutes only 3 per cent. Quartz is over twice as abundant as orthoclase and exists in two forms—as very small well-formed doubly terminated water-clear crystals locally resorbed and as a secondary rim of turbid quartz in optical continuity. In some sections these rims are composed of chalcedony. The crystals attain a maximum size of 0.2 mm. Orthoclase may occur as idiomorphic, clear crystals without rims or turbid irregular grains mixed with quartz. Abundant quartz, sericite, carbonate, and pyrite have been introduced hydrothermally.

## CONTACT METAMORPHISM

Of the three areas where contact metamorphism has been operative the largest completely bounds the exposed margins of the monzonite stock. A limited area is noted near the Oro Plata Mine on the northwest side only of the Oro Plata intrusion, while only a small patch of affected rock is to be found on the western side of the Orient granite porphyry. The wall rocks of the dikes and smaller intrusives outside of these instances show no extensive change other than occasional pyritization and slight recrystallization of the carbonate minerals. One thin section of a specimen taken from within 2 inches of an aplite dike in the Old Scotia Mine shows incipient silicification.

The contact zone of the monzonite attains its greatest width—2500 feet—directly east of the Tintic Western Mine and narrows toward the north and the west to about 750 feet.

The development of metamorphic minerals seems to depend on two factors, proximity to the intrusion and original composition of the sediment. Many rock types are present, therefore, and several of these will be described.

Near the Tintic Western Mine, almost at the contact of the monzonite, a coarsely crystalline calcite marble has prominent laminations or streaks

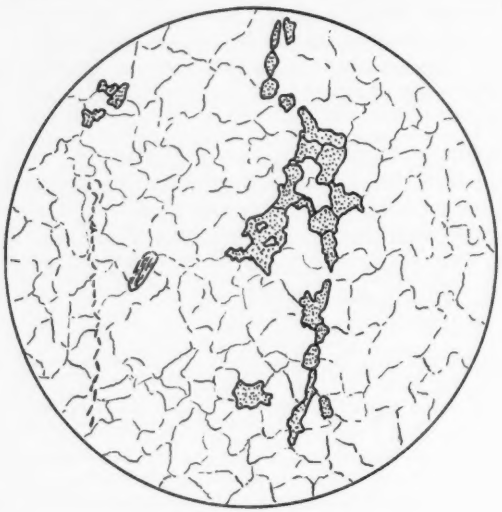

FIGURE 2.—*Orient quartzite*
Trains of apatite and sericite in a mosaic of quartz. X 20.

composed of bottle-green idocrase alternating with white fluorescent wollastonite. These laminae are remarkably constant along the strike of the beds but may alternate across the strike with wide zones of pure blue calcite (Pl. 2, fig. 2). Under the microscope bands of quartz, orthoclase, and diopside parallel and alternate with these laminae. Scattered throughout the entire rock and included in all minerals are small epidote crystals. This bed apparently was a dolomite with many layers of argillaceous and arenaceous material which furnished the alumina, silica, and potash for the development of the silicates.

Near the Iron King Mine is a massive and slightly green bed. The microscope shows chiefly fine carbonate with bands of serpentine in closely packed rounded patches evidently derived from previously existing forsterite. East of Bates Gulch a dense massive gray rock shows idocrase, quartz, wollastonite, orthoclase, birefringent grossularite, diop-

side, and coarse calcite. The Orient shale has been changed to a dense, dark-brown hornfels, consisting of very small quartz grains and minute mica flakes intimately intergrown with occasional patches of felted sericite. Many limonite cubes evidence the former extensive development of pyrite. The Orient quartzite is composed predominantly of interlocking grains of clear quartz. Trains of small stubby crystals of apatite and flaky sericite which are irregularly distributed suggest introduction of these minerals (Fig. 2).

A few feet east of the Orient quartzite, radiating aggregates of tremolite appear in milk-white marble in the form of solid lenses 3 inches thick which weather out as knobs. A very little microscopic diopside is mixed with the tremolite and is of early development. This is the easternmost outcrop containing diopside.

Eastward up Brown's Ridge, large areas of white marble with or without tremolite are observed. Near the margin of the affected area beds of blue dolomite alternate with the marble and may or may not contain tremolite. A specimen was found in which tremolite had developed in blue dolomite containing *Halysites*. In one prospect on the east side of Brown's Ridge there has been extensive development of tremolite in the blue limestone. Here quartz and a little galena are associated with it.

Along the strike northward from the vicinity of the Iron King Mine, minerals recognizable in hand specimen become smaller. Near the northern margin of the contact area only a few specimens containing tremolite were found. Farther west Proterozoic quartzite and phyllite are extensively recrystallized with development of considerable biotite.

An attempt was made to discover if the smaller intrusions in the marmorized area had exerted any extra influence in the metamorphism, but none could be detected.

In the vicinity of the Oro Plata mine exposures are poor, and no gradation in intensity of metamorphism was evident. However, specimens of dense dark-gray dolomite with black bedding laminae were collected. These are composed principally of a mosaic of carbonate crystals, but the black bands contain chondrodite, phlogopite, spinel, graphite, and serpentine altered from the chondrodite.

The contact action of the Orient intrusions may have been more extensive than indicated since it is almost entirely surrounded by quartzite which would be very little changed. Slate, however, is soft and brown, and under the microscope it was found that quartz and biotite show incipient recrystallization, and pyrite now altered to limonite has been introduced.

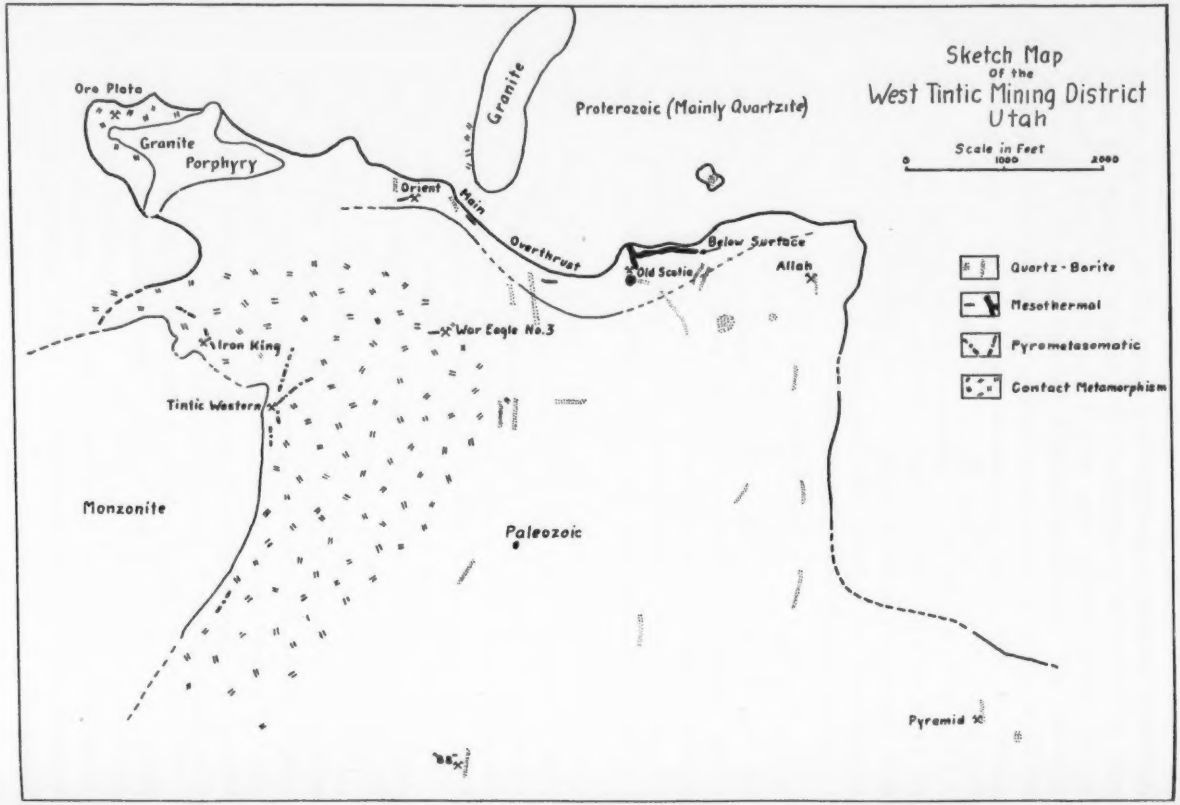


FIGURE 3.—*Sketch map*
Shows the distribution of quartz-barite, mesothermal, pyrometasomatic deposits, and contact metamorphism.

## MINERAL DEPOSITS

### PYROMETASOMATIC VEINS[1]

Veins of the pyrometasomatic type are distributed within an area 600 feet from the contact of the monzonite. Most of them are situated at the north and eastern borders, but a few minor ones can be found toward the south (Fig. 3). The Tintic Western Mine afforded the only underground observation. The veins vary in width from 3 and 4 feet to a fraction of an inch (Pl. 2, fig. 3). It was found that they traversed monzonite porphyry, granite porphyry, and contact metamorphic limestone. Individual crystals in vugs may attain a maximum size of 2 or 3 inches and grade to microscopic dimensions.

Grossularite, clinozoisite, andradite, quartz, chlorite, epidote, calcite, actinolite, scheelite, magnetite, specularite, chalcopyrite, and pyrite were identified. Stibnite and lead minerals have been reported. These minerals do not all occur together in one place, but it is believed that they are of nearly contemporary formation and are all part of the same general episode of mineralization. In a prospect near the Iron King Mine, cream-colored coarsely crystalline grossularite occurs with light-tan clinozoisite in radiating aggregates together with a little quartz and calcite. Andradite is associated with all other minerals except grossularite. The specific gravity was determined upon several specimens of these two garnets with the following result:

| Andradite: | Grossularite: |
|---|---|
| 3.829 | 3.622 |
| 3.841 | 3.641 |
| 3.835 | 3.642 |

The variation of these values is probably due to impurities. However, they all fall well within the specific gravity limits of these two species as given by Dana (1892, p. 440) and are hence regarded as diagnostic. Near some veins andradite has penetrated the marble in fine veinlets or developed in isolated crystals (Pl. 2, fig. 3). At the Tintic Western Mine andradite is associated with white quartz, epidote, fine-grained brownish quartz, chalcopyrite, platy magnetite, fibrous actinolite, chlorite, calcite, and scheelite. In a prospect west of the Iron King Mine epidote is particularly abundant in a vein of calcite, quartz, scheelite, and andradite. The vein exposed at the Iron King is composed principally of specularite and magnetite with some andradite and quartz, with pyrite veinlets present in each of these minerals. Chalcopyrite is present in small amounts, both at the Tintic Western and farther south in veins with quartz and garnet. Some of these are

FIGURE 3.—*Sketch map* Shows the distribution of quartz-barite, mesothermal, pyrometasomatic deposits, and contact metamorphism.

[1] This classification follows Wandke and Moore (1935, p. 780) who use the term for deposits which are not derived from the immediately adjacent igneous rocks but have risen from depth from a deep-seated magma perhaps unexposed.

so small as to be measured in millimeters and are often repeated in a single hand specimen, imparting to it a striped appearance. Fine-grained quartz with brownish stains deposited on coarse white quartz is associated with chlorite, scheelite, and pyrite. Coarse crystalline quartz with brownish surface stains is found in one long vein, essentially

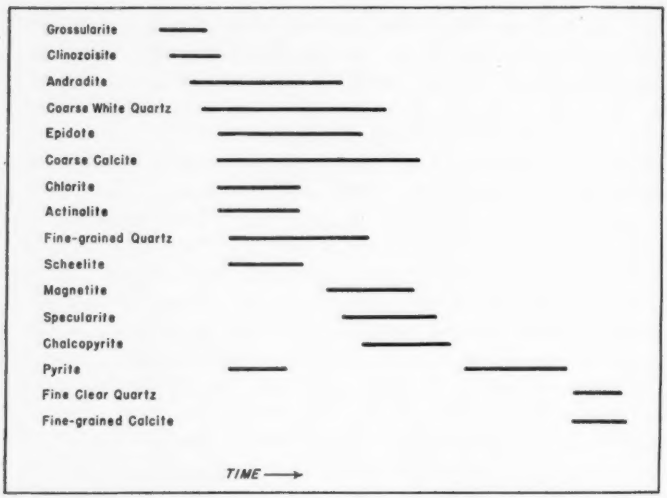


FIGURE 4.—*Sequence of mineral formation in the pyrometasomatic veins*

without appreciable amounts of other minerals. Veinlets of fine clear quartz and granular calcite traverse all minerals.

In considering the various occurrences as a whole, a paragenesis has been worked out (Fig. 4). Oxidation of the chalcopyrite has resulted in the development of malachite, chrysocolla, and a black copper pitch mixed with limonite. Limonite, partly pseudomorphous after pyrite, is present throughout all the deposits.

### MESOTHERMAL VEINS AND REPLACEMENTS

This type of deposit is limited to an area very close to the main overthrust between the Orient and Old Scotia mines (Fig. 3). An exception is one small vein on the western slope of Brown's Ridge at a prospect called War Eagle No. 3. The conditions of the Orient workings prevented observation underground, but one 3-foot vein is exposed for several yards at the surface, dipping steeply to the north. It is composed of oxidized ore containing a high concentration of cerussite and anglesite.

The War Eagle No. 3 vein trends about N. 75 E. and dips steeply northward and is situated just within the border of the marbleized area. The only primary minerals noted at the surface here were a little galena, pyrite which has been oxidized to a reddish earthy mass of limonite and hematite, and small amounts of cerussite and anglesite. Gold values have been reported. Two prospects east of the Orient Mine show a little galena, pyrite, and accompanying oxides.

The principal studies of this type of mineralization were made at the Old Scotia Mine where good exposures are available both at the

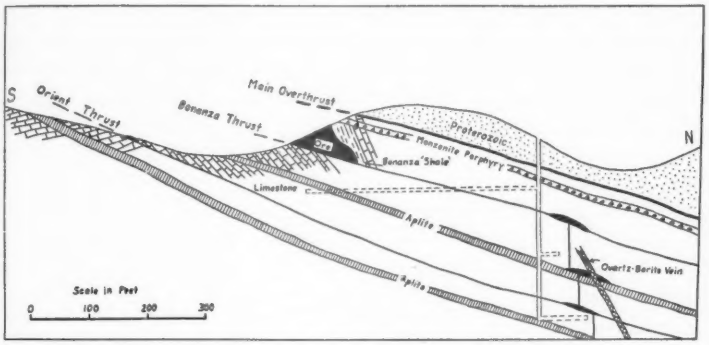


FIGURE 5.—*Generalized section through the Old Scotia mine*
Shows the position of the mesothermal ore bodies in relation to the aplite dikes and thrust blocks. Ore bodies are elongated perpendicular to the section.

surface and underground. The key to the structural control of the emplacement of the ore here is found in the discovery in Glory Hole where the ore is localized in the brecciated zone of the Bonanza overthrust (Fig. 5). The ore bodies have the form of nearly flat-lying blanket replacements, with very irregular local distribution but with a general east-west elongation. Many small irregular pipes branch from these. It can be demonstrated that some blanket ore shoots lie along flat-lying faults. In other instances, 3- or 4-foot seams have been mined from directly above aplite, where apparently these dikes have determined to a large extent the position of the ore (Fig. 5). Small, vertical, east-west veins were found in the floor and ceiling of several stopes. Of particular interest is the vertical Blue Jay vein, composed of quartz and galena with abundant malachite stains, which strikes a little north of east and is the most constant and predictable feature in the mine. It forms approximately the northern limit of one of the larger ore

bodies which lies on top of aplite. In viewing the physical relations of the mineralization broadly, at the Old Scotia, it appears that the ore-forming solutions ascended in nearly east-west vertical fissures and where they encountered a brecciated zone, resulting from a flat-lying thrust fault or an aplite dike, replaced the limestone, forming an ore body.

The ore is so completely oxidized, even at a depth of 300 feet, that only a little of the primary ore is left. This is composed principally of galena, arsenopyrite, and pyrite, with minor amounts of sphalerite and famatinite and a very little chalcopyrite. Polished specimens of these minerals show that they have been deposited in the following sequence: arsenopyrite, pyrite, galena, sphalerite, and famatinite (Fig. 6). Gold and silver minerals could not be detected, although approximately half of the $40 per ton average is in these metals. Quartz is the principal primary gangue mineral but is present only in subordinate amounts, while in the Glory Hole a little microscopic adularia occurs.

Ore-forming solutions have affected the country rock very little where it is composed of pure dolomite or limestone. Where chert is present, however, talc has been formed. In the Glory Hole the Bonanza formation has yielded much talc and chlorite, probably as the result of the very impure dolomitic character of this rock. As has been mentioned previously, the monzonite porphyry is extensively altered to sericite and carbonate.

Underground the sulphides are oxidized to a deep red hematite mixed with cerussite and anglesite. Scorodite, particularly abundant near sulphide nodules, has developed from arsenopyrite early in the oxidizing process, but as oxidation progressed it changed to hematite. Circulating waters have deposited gypsum in the vugs of the sulphide. In the Glory Hole, at the surface, hematite is subordinate to limonite, and a great many other oxides are seen. Conichalcite is abundant as light-green botryoidal crusts. Olivinite, malachite, chrysocolla, azurite, linarite, and a canary-yellow mineral in powdery crusts, thought to be bindheimite, are also present.

Late fissures have been filled with travertine. These traverse ore and country rock alike and are entirely unrelated to the primary mineralization.

### QUARTZ-BARITE DEPOSITS

Quartz-barite veins, generally carrying a little galena, are so abundant throughout the eastern two-thirds of the area that only the larger and more important ones could be represented on a map (Fig. 3). It is significant, however, that in the areas of contact metamorphism only one or two small stringers were found. In only one deposit, east

of Brown's Ridge, tremolite, quartz, barite, and a little galena occur together, in veins and replacement bodies. Of these, the former is most abundant.

The veins range from a fraction of an inch to 3 or 4 feet in width, and a single outcrop may vary from one extreme to the other within a distance of 10 feet along its strike. Small veins may branch from one another or grade into a wide zone of closely interlacing veinlets which produce a magnificent boxwork structure when weathered. Vugs are common, and quartz crystals approaching half an inch in length are observed. The quartz is usually white with brownish stains, and occasionally opaque white plates of barite are recognizable. Galena is present in many instances, varying from microscopic to quarter-inch cubes. The presence of this material undoubtedly promoted most of the early prospecting. Every known large quartz-barite vein carrying galena shows prospecting. Several deep workings have been made on these veins, notably the Allah, Pyramid, and "88" mines. Except for a very few instances where galena was concentrated in small pockets, no large shipments have been made to date. Very rarely the quartz is stained with malachite, and one grain of chalcopyrite was found. A little sphalerite was reported from the Allah vein. At the "88" Mine considerable clear to purple fluorite is associated with quartz and galena. Cerussite, anglesite, and chrysocolla are present in microscopic amounts.

The replacement bodies are well exemplified by a large oval outcrop 50 feet in smallest dimension, exposed on the nose of a southward-sloping hill southeast of the Old Scotia Mine near the fork of the road. The quartz is brown and very dense, resembling certain of the more massive types of Proterozoic quartzite. Vugs and stringers are absent. Galena is present as small dull crystals. Microscopic study reveals that barite is very abundant as plates and irregular grains. Relict structures of the brecciated limestone are also evident. In the Old Scotia Mine a quartz vein cuts the main ore at one place (Fig. 5), while several instances of replacement quartz can be found. Thin-section study showed that there was no regular sequence of deposition of quartz, barite, and galena, suggesting essential contemporaneity.

## SEQUENCE OF EVENTS

The sequence of igneous intrusion and periods of mineral development, judging from field evidence, is fairly clear. The monzonite and monzonite porphyry are not found together, and their relationships were not observed, but from the close similarity of the two rocks it is believed that they are from the same magma and practically contemporaneous. They were also certainly intruded after the episode of folding and fault-

ing, since the monzonite cuts the main overthrust and the position of the monzonite porphyry dikes has been controlled in many instances by the fracturing resulting from the thrust.

Granite porphyry intrudes the monzonite in two places along its northern border (Pl. 1) and is therefore regarded as later. It has already

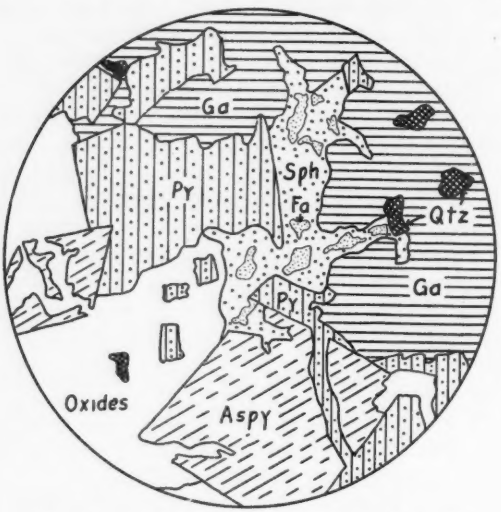


FIGURE 6.—*Sulphides from the Old Scotia mine*

Arsenopyrite is traversed by a veinlet of pyrite. Galena replaces pyrite, and sphalerite and famatinite replace galena, pyrite, and arsenopyrite.

been shown that the granite is later than granite porphyry, and an aplite dike was found within the granite mass which offers proof of its later intrusion.

The contact metamorphism has not such a definite position in the true sequence since neither aplite nor granite is found in contact with rocks affected by this process. The effects adjacent to the Oro Plata and Orient bodies must follow these intrusions. The large area near the monzonite could have developed directly after the intrusion of that body, but since no additional effect was found resulting from the intrusion of the granite porphyry in this area it is thought that all the contact action came later. Further intrusion of granite and aplite followed the granite porphyry, and the magma reservoir was therefore still charged with rock-forming melt, and volatiles were not yet concentrated enough to be expelled and accomplish the metamorphism. It is consequently reason-

able to assume that the entire contact action took place after the intrusion of aplite.

The pyrometasomatic veins cut contact rock and are therefore later. The relationship of the mesothermal deposits to the pyrometasomatic veins is not known since no contacts between the two are found, but the presence of the small War Eagle No. 3 vein in tremolite marble establishes its age as later than the contact action. Aplite dikes have controlled the emplacement of some of the ore in the Old Scotia Mine (Fig. 5) which proves quite definitely that these bodies are also later than the dikes.

The quartz-barite mineralization traverses the ore at the Old Scotia Mine (Fig. 5), and many instances of quartz veins in aplite and granite porphyry were found. These deposits may safely be considered therefore to have developed later than the aplite and mesothermal deposits. Figure 7 represents the order of the principal events involved in the magmatic and mineralogic history of the district.

## CONDITIONS DURING INTRUSION AND MINERALIZATION

The monzonite stock furnishes evidence that the first magma to rise from depth was slightly oversaturated with silica. The outcrop, though present only in the southwest, may be simply the top of a cupola which expands downward until it eventually underlies the entire district, or it may be the top of a long narrow vertical or inclined cylinderlike plug or the top of a funnel-shaped body. In either case it may be supposed, because of the greater evidence of contact effects, that this exposure represents the approximate center of the main igneous activity in the district. During this intrusion some of the monzonitic magma escaped and made its way along fractures and fissures in the country rock forming the monzonite porphyry dikes. Activity thereafter ceased for a time and allowed at least the exposed portion to solidify. Contemporaneous with this lull, differentiation progressed to the point where the magma when it next appeared at the present erosion level was of granitic composition and resulted in the formation of granite porphyry. This invasion, though less in quantity, was not so restricted in area as before. More numerous but smaller bodies were formed in monzonite and sedimentary rocks alike. Another surge of magma brought in a granite which in three instances followed the old channels of the granite porphyry. The last stage of differentiation produced a highly siliceous residue which formed dikes of aplite.

This entire sequence of intrusion appears to represent the normal succession of events expected to occur during the differentiation and

cooling of the upper portions of a magma body. The composition of the original magma of course cannot be postulated, but following Bowen's (1933) theory of fractional crystallization the first liquids should be more basic than the last, and aplites are usually thought to be formed at lower temperatures.

The conditions which might be postulated to prevail immediately after the aplite intrusions are as follows: Temperatures were still high but not

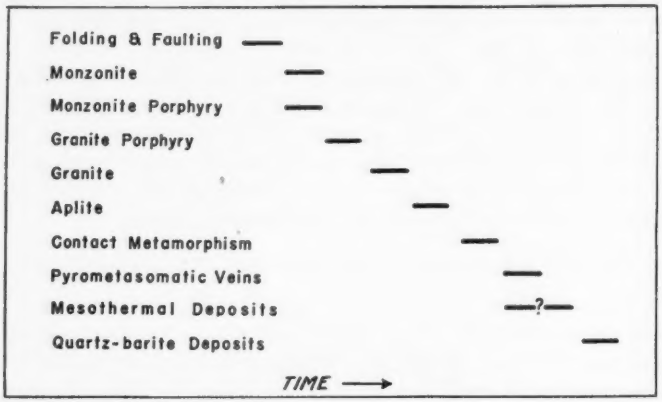


FIGURE 7.—*Sequence of igneous intrusion and periods of mineralization*

uniform; the cooler portions were those at considerable distance from the larger intrusive bodies; the progress of differentiation in the magma had reached the point where a complex silicate melt was no longer available and a liquid containing a large amount of volatile constituents had developed. This material found its way into heated areas surrounding the igneous bodies and effected the contact metamorphism. Evidence for two stages of contact action as found by Umpleby (1917, p. 65) at Mackay, Idaho, is lacking, but marmorization may have preceded the development of the silicates, although the formation of the latter has masked entirely the earlier action. It is thought, therefore, that the major part of the contact metamorphism was accomplished after the solidification of the intrusive magma. This hypothesis follows Prescott (1915, p. 68) who concludes that "at least to the greatest depth reached the intrusion is often a solidified mass capable of fracture at the time of formation of the contact silicates." It has been shown that the growth of the contact silicates was accomplished largely from the constituents already present in the sedimentary rocks. Therefore the metamorphosing liquid was apparently extremely rich in water and poor in

other elements. This may also be substantiated by the fact that only two of the usual four contact metamorphic zones postulated by Hess and Larsen (1921, p. 253), the light silicate and marble zones, are present. This would suggest that, although the total width of the two zones is great, the solutions which were abundant were unusually weak in silicate-forming materials. Temperatures must have been fairly high near the monzonite where wollastonite and grossularite have developed and became less with distance where tremolite is found in greatest abundance, and even the quantity of this mineral diminishes rapidly toward the periphery of the contact zone. With further cooling water was still present in the rock as evidenced by the serpentinization of chondrodite and forsterite. The presence of chondrodite and apatite suggests that fluorine accompanied the solutions and took part in the development of these minerals.

Differentiation in the magma was still progressing, and silica, iron, copper, lead, arsenic, antimony, tungsten, gold, silver, and zinc began to be expelled along with volatiles. Near the monzonite where temperatures were still high the solutions deposited the pyrometasomatic veins. The source of aluminum, calcium, and magnesium contained in the minerals of these veins is unknown. They may have been present in the original differentiated liquid or may have been picked up from the igneous and sedimentary rocks enroute. In the cooler areas where fracturing encouraged extensive depositions, these solutions deposited arsenopyrite, pyrite, galena, sphalerite, famatinite, a little quartz, and adularia, in an environment usually attributed to mesothermal conditions. The mineral sequence in both these deposits indicates progressive cooling during deposition. Grossularite and other high-temperature silicates together with scheelite were the first to be deposited in the pyrometasomatic veins, followed by oxides, sulphides, and fine crystalline quartz and calcite. In the mesothermal deposits arsenopyrite, usually considered to be a high-temperature mineral, is followed by other sulphides. Adularia, a low-temperature mineral, though not found in direct association with sulphides, is believed to represent the final stages of hydrothermal deposition. Alteration was accomplished by these solutions, and talc, chlorite, sericite, and carbonate were developed in the country rock adjacent to the ore bodies.

A later influx of mineralizing solutions, carrying a great amount of quartz, barite, and a little galena, also apparently derived from the still active end-stage processes of the magma, occurred next. Temperatures were still too high in the contact zones, except in one place at the margin, to allow deposition of this material, and it was carried beyond

to cooler regions, forming veins and replacements where conditions were more favorable.

## SUMMARY OF CONCLUSIONS

The field and mineralogical data on intrusion and hydrothermal deposition in the West Tintic Mining District furnish evidence which closely links these two processes. A series of intrusions of successively more silica-rich rocks ending in aplite is followed by contact action and two stages of hydrothermal deposition. At least three stages of escape of materials from the magma after its consolidation are registered: (1) a liquid very rich in water was expelled and accomplished contact metamorphism in rocks still highly heated by preceding magmatic invasions; (2) very shortly thereafter solutions rich in silica, iron, copper, lead, arsenic, antimony, tungsten, gold, silver, and zinc found their way into the country rock. Near the heated contact, pyrometasomatic veins were developed, while in the cooler regions minerals were deposited at medium (mesothermal) temperatures; (3) later, solutions rich in silica, barium, and sulphur and poor in lead appeared. The contact regions and the larger intrusives were still too hot to allow this assemblage of elements to be deposited, and they migrated to cooler regions where deposition was effected in sedimentary rock, smaller intrusives, and already cooled mesothermal bodies. The whole succession of events, therefore, is believed to represent a history of cooling magma, differentiation, and progressively lower-temperature deposition of hydrothermal minerals.

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# FOLDED THRUST FAULTS IN THE FOOTHILLS OF ALBERTA

BY B. F. HAKE, ROBIN WILLIS, AND C. C. ADDISON

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## ABSTRACT

This paper describes the stratigraphic and geologic structure of an area in the Rocky Mountain foothills of west-central Alberta. An unusual group of low-angle thrust faults is described and analyzed. At least seven overlapping faults were recognized which range up to 30 miles in length and have an aggregate displacement of about 6 miles. They divide the incompetent Upper Cretaceous sediments into thin sheets. The faults and faulted sediments have been strongly folded. These faults are thought to have developed in an asymmetrical syncline where the structural development required more bedding-plane slippage than the strata were capable of performing and the faulting relieved the stresses involved. The faults are believed to be confined to the stratigraphic section and the major structure with which they are found associated and to die out completely with depth without producing any dislocation of Paleozoic rocks.


## INTRODUCTION

### SCOPE OF PAPER

In this paper an unusual group of low-angle thrust faults is described and analyzed. The area herein dealt with contains at least seven such overlapping faults, which range up to 30 miles in length and have an aggregate displacement of about 6 miles. They divide the incompetent Upper Cretaceous sediments into thin but extensive sheets. These and the fault planes have been strongly folded.

This group of faults is considered noteworthy for the following reasons: (1) They are developed in weak beds, and the fault planes lie at an angle to the bedding, so that the sheets themselves are not competent to have transmitted the thrust which caused the displacement. (2) The faults bear an exceptionally systematic relation to the bedding, and the

folded planes are repeatedly well exposed, so that not only actual displacements, but variations therein can be determined.

It is believed that these faults developed in an asymmetrical syncline requiring an amount of bedding-plane slippage which the in-

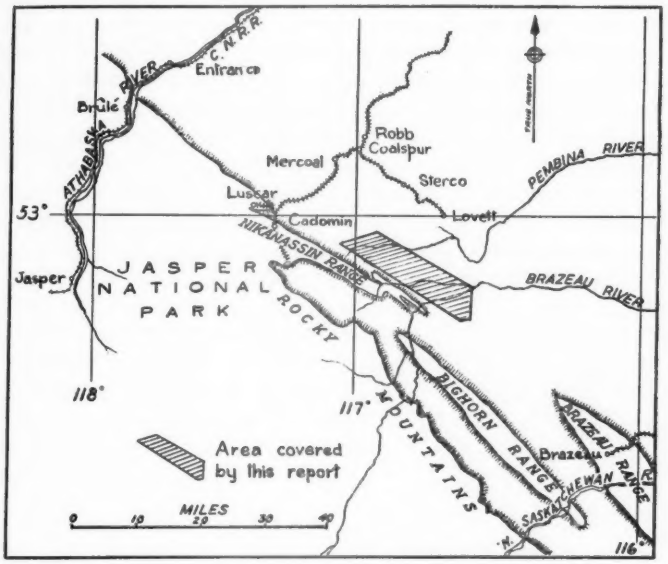


FIGURE 1.—*Index map*

Showing location of Cardinal district and adjacent mountain ranges.

volved strata were incapable of performing. The faulting and the attendant crumpling relieved stresses which in other folds and in other stratigraphic sections are relieved by bedding-plane slippage.

In spite of their considerable magnitude these faults are thought to be confined to the stratigraphic section and the major structure with which they are associated, and to die out completely with depth without producing any dislocation of the Paleozoic rocks of the district.

## FIELD WORK

Part of the drainage area of the Brazeau, Cardinal, and Pembina rivers in the Rocky Mountain foothills of Alberta, east of Jasper Park (Fig. 1) is here described and referred to as the Cardinal district.

The field work for this report was done during the summer of 1929; the senior author also worked in the area during part of the summer

of 1928, in company with J. B. Webb, who had spent previous months working on the stratigraphy and structure of this and other foothills areas. Mr. Webb's excellent stratigraphic work of 1927 and 1928 laid the foundation on which the authors were able to work out a sufficiently detailed stratigraphic column to enable them to interpret the structural complexities observed.

Pace traverses were carefully made along the rivers and were supplemented by a great deal of work in the interstream areas with the aid of topographic maps published by the Geological Survey of Canada. Considerable additional work could have been done in checking the continuity and correlation of faults and folds from stream to stream had time permitted. A complete solution of the structural problems in this area could only be made after a great deal of drilling had been done, but the writers believe that their work demonstrates the type of structure present and shows some of its significant details.

The field data are too complex and voluminous for detailed reproduction with this paper. Most of the traverses were plotted in the field on the scale of 1000 feet to the inch and even on that scale it was difficult to find room for legible recording of all the dips and stratigraphic contacts observed and used in the interpretation. In order to present an intelligible picture of the conditions believed to exist in this district, it has been necessary to connect apparently related outcrops and structural features across areas of no outcrop. On any scale capable of satisfactory reproduction herewith, it would be impossible to indicate the exact extent to which this has been done. Observations in several areas where the structural details were more clearly exposed are shown in some of the accompanying figures.

### PREVIOUS WORKERS

The subdivisions of the marine Upper Cretaceous formations herein presented are in substantial agreement with those noted previously by Warren and Rutherford (1928), but the writers have further subdivided the formations of this area and have established the thicknesses and limits of the units with greater precision.

Various features of the area are mentioned in the reports of reconnaissance surveys by McEvoy (1898), Dowling (1909), and Stewart (1916). Geological reports and maps of the area were published by Allan and Rutherford (1924) and Rutherford (1925); MacKay (1930) also has published geologic structure sections across parts of this area. However, none of these workers appears to have recognized the low-angle thrusts here described.

## TOPOGRAPHY

The Cardinal area exhibits moderately rugged mature topography with a maximum relief of more than 1000 feet and a local relief of 500 to 800 feet. The average elevation is between 5000 and 5500 feet above sea level.

The topography is in general related to geologic structure. Strike ridges of the more resistant sandstones are paralleled by broad valleys carved in the softer shales. The principal streams are transverse, and the longitudinal streams comparatively short, the longest being Crooked Creek which extends for 5 miles along a strike valley. The drainage system developed on a gravel plain formed during the early or middle Tertiary, and has been superimposed on the Cretaceous rocks, so that the present day streams are only partially adjusted to the structure. The principal streams are entrenched below broad valleys. Along their courses and on some ridges the rocks are excellently exposed. Elsewhere, outcrops are neither numerous nor extensive.

For several miles above the confluence of the Brazeau and Cardinal rivers, moraines conceal the Cretaceous rocks, give the surface a very rugged character, and form several small lakes. A few small meadows exist in the valley bottoms, but most of the area is covered either by pine and spruce timber or by swamps of the type known as "muskegs."

Southeast and southwest of this area the Brazeau and Bighorn ranges (Fig. 1) rise to rugged peaks of 7000 to 8000 feet elevation. West of the area the Brazeau River flows eastward through a broad pass between the northern end of the Bighorn Range and the southern end of the Nikanassin Range. This latter is a spur which increases in elevation toward the northwest and merges with the main Rocky Mountain front near the Athabaska River.

## STRATIGRAPHY

### SUBSURFACE FORMATIONS

Although rocks older than the Colorado series of the Cretaceous are not exposed in this area, they are present in considerable thickness at depth and are important to structural theory.

MacKay's (1930, p. 3-10) work on the earlier sediments was much more detailed than that done by the writers and his reported thicknesses are therefore quoted. The senior author has examined numerous sections of these older rocks in adjacent areas and his observations are in general accord with MacKay's. However, the thicknesses of these rocks beneath this area may be somewhat less as the Mesozoic section exposed in the Brazeau Range appears to be thinner than in the Nikanassin Range.

SUBSURFACE FORMATIONS OF THE BRAZEAU REGION
(after MacKay)

| Age | Formation | Description | Thickness (in feet) |
|---|---|---|---|
| Lower Cretaceous | Mountain Park | Brown sandstones and green shales with lenses of conglomerate | 400 |
| | Luscar | Gray sandstones and shales with coal measures | 1,700 |
| | Cadomin | Massive conglomerate | 35 |
| | Nikanassin | Gray shales and sandstones with thin coals. | 1,900 |
| Jurassic | Fernie | Black marine shales and thin limestones | 1,300 |
| Triassic | Spray River | Dark-gray limestones and shales | 700 |
| Pennsylvanian | | Quartzite | 90 |
| Mississippian | | Chiefly massive limestone | 1,500 |
| Devonian | | Chiefly massive limestone | 3,300 |
| Cambrian | | Chiefly massive limestone | 2,000 |
| | | Total | 12,925 |

## SURFACE FORMATIONS

The surface rocks involved in the low-angle thrusting are all Upper Cretaceous. It is not known whether the time-equivalent of the Dakota sandstone is a portion of the Blackstone formation, or of some of the rocks here classed as Lower Cretaceous. Table 1 shows the Upper Cretaceous column of this area and equivalent formations in other areas of the foothills.

TABLE 1.—*Upper Cretaceous Rocks of Alberta Foothills*

| Bighorn Basin [1] | Saunders Creek [2] | Highwood River [3] | Turner Valley [4] | Blairmore Area [5] |
|---|---|---|---|---|
| Brazeau | Lower Saunders formation | Belly River | Belly River | Allison |
| Wapiabi | Wapiabi | Upper Alberta Shale | Upper Benton Shale | |
| Bighorn | Bighorn | Cardium Bands | Lineham Sandstone member | Benton Niobrara |
| Blackstone | Blackstone | Lower Alberta Shale | Lower Benton Shale | |

[1] Malloch (1911).
[2] Allan and Rutherford (1923).
[3] Hume (1929).
[4] Slipper (1921).
[5] Leach (1912).

On account of their priority and fitness the names given these formations by Malloch (1911) are preferred by the writers. The thicknesses given in the following section are in some cases averages of those observed by the authors, and in others those believed to be most nearly representative of general conditions in the area.

UPPER CRETACEOUS ROCKS EXPOSED IN THE BRAZEAU REGION

| | Thickness in feet |
|---|---|
| BRAZEAU FORMATION: | |
| Heavy-bedded continental deposits of sandstone and greenish-gray shale with carbonaceous matter and coal | 5000+ |
| WAPIABI FORMATION: | |
| (5) Blocky zone: massive, somewhat concretionary gray sandy shale with abundant *Baculites ovatus* | 250 |
| (4) Upper concretionary zone: highly concretionary, rather sandy gray to rusty shale | 240 |
| (3) Platy shale zone: thin-bedded to laminated gray shale and fine sandstone with small fossils, (notably *Ostrea aff. congesta* and *Baculites* cf. *anceps*) occurring in calcareous lentils | 440 |
| (2) Lower concretionary zone: massive to fissile rusty shale with very numerous concretions of widely diverse sizes; characterized by abundance of *Scaphites ventricosus* | 550 |
| (1) Striped zone: shale and sandy shale with rare concretions, weathering to a rusty banded appearance | 120 |
| Total | 1600 |
| BIGHORN FORMATION: | |
| (6) Upper sandstone: thin-bedded, hard flaggy white to yellow sandstone with minor shale; characterized by *Cardium pauperculum* | 60 |
| (5) Upper shale: massive sandy shale with prominent ironstone bands | 70 |
| (4) Middle sandstone: massive muddy siltstone, more resistant than the enclosing shales | 20 |
| (3) Lower shale: sandy concretionary shale similar to upper shale but with notably less ironstone | 75 |
| (2) Lower sandstone: clean white moderately thick-bedded sandstone frequently carrying abundant small rugose species of *Inoceramus* and with local coal at top of member | 50 |
| (1) Transition zone: flaggy sandstones and thin shales; sands becoming thinner and less frequent downward | 150 |
| Total | 425 |
| BLACKSTONE FORMATION: | |
| (3) Hydrogen sulphide zone: massive black shale with few thin laminae of sand and a few thin persistent bentonite layers; weathers very rusty and usually smells of hydrogen sulphide | 300 |
| (2) *Inoceramus labiatus* zone: black fissile shale with sandy laminae, characterized by presence of *Inoceramus labiatus* and *Prionotropis* sp. | 350(?) |
| (1) Barren zone: dark-gray to rusty well-bedded sandy shale without notable fossils | 250(?) |
| Total | 900(?)* |

* The lower part of the Blackstone formation is not so definitely separable nor so well exposed as the higher formation members and consequently the thicknesses of the two lower zones are of questionable accuracy.

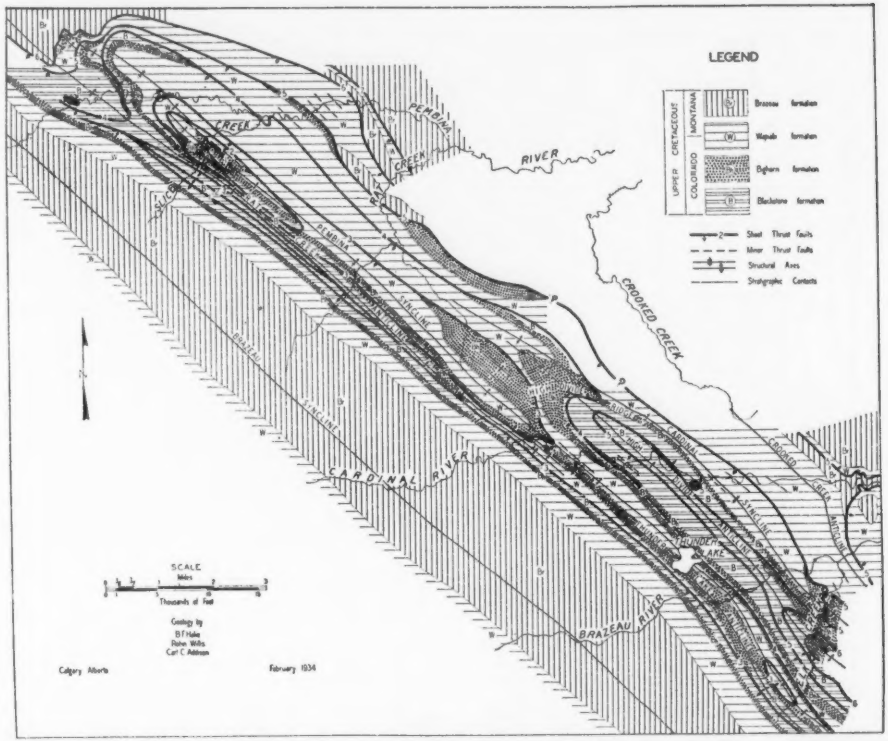


FIGURE 2.—*Geologic map of Cardinal district*

Only about 5000 feet of the sediments belonging to the Brazeau formation are present in the Brazeau syncline in the Cardinal district. Stratigraphic studies (1926) farther eastward have shown that above this lower portion of the Brazeau formation 7000 to 9000 feet of younger sediments was deposited. Thus at the beginning of the late Cretaceous-early Tertiary Rocky Mountain orogeny the Paleozoic limestones in the Cardinal district may have been overlain by Mesozoic and Tertiary sediments to a total thickness of between 20,000 and 25,000 feet.

## REGIONAL STRUCTURE

The Cardinal district is the northwestern half of a portion of the foothills belt which is partially enclosed by spur and outlier ranges of the Rocky Mountain system. The Brazeau Range, to the southeast, is an anticline plunging in both directions from its central section and exposing Devonian rocks in its core. Its eastern flank is faulted so that Paleozoic limestones rest upon the upturned edges of Mesozoic strata. The Bighorn Range to the southwest is also a faulted anticline, as is likewise the Nikanassin Range west of the Cardinal district. To the northeast the sediments are gently folded but dip generally toward the trough of the Alberta syncline.

The Cardinal district has two principal structural units: a syncline, separated from the western ranges by a zone of tight folding and thrust faulting, and to the east of the syncline, a long narrow anticlinorium with which this paper primarily deals.

The syncline contains strata of the Brazeau formation and is here called the Brazeau syncline (Fig. 2). It extends throughout the length and beyond the area, plunging gently southeastward. On its northeastern flank it presents a fairly uniform series of dips ranging from 10° to 60°; the southwestern limb is much steeper and at the contact between the Brazeau formation and underlying Wapiabi, the dips range from 80° NE., to 70° overturned SW.

The principal structural features of the anticlinorium are four anticlines and two synclines. The structural elements on the Brazeau River from southwest to northeast are referred to as the Thunder Lake anticline, the Pembina syncline, the High Divide anticline, the Cardinal syncline, and the Crooked Creek anticline. In the Pembina river section the Crooked Creek anticline was not recognized. The High Divide anticline and Cardinal syncline seem to have lost their identities in the eastern limb of the Pembina syncline. The Rat Creek anticline immediately southwest of the Pembina syncline is almost but not quite continuous with the Thunder Lake anticline. The structure of

FIGURE 2.—*Geologic map of Cardinal district*

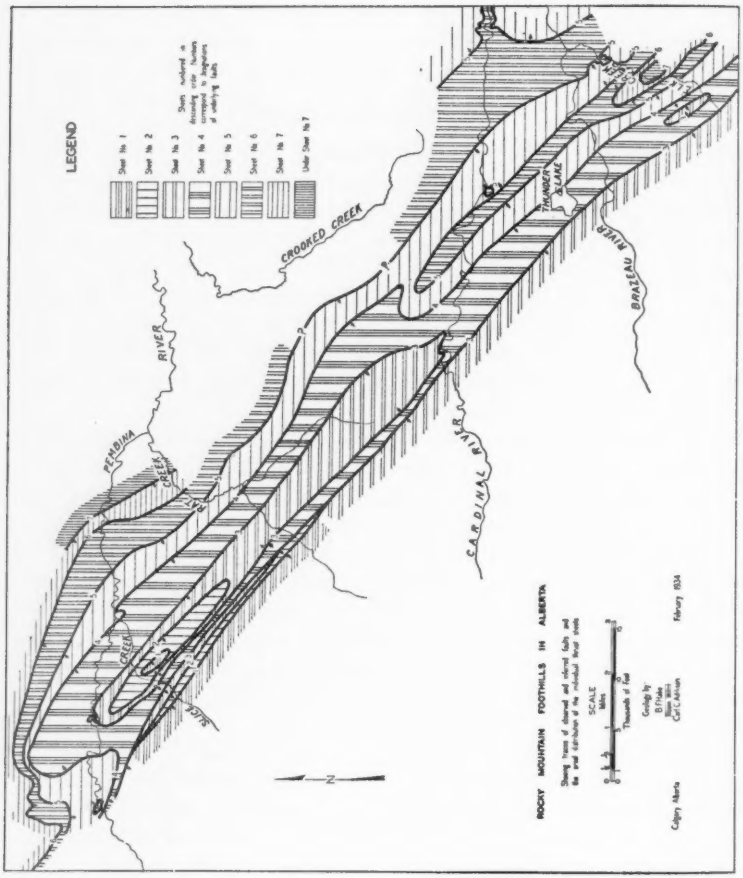


FIGURE 3.—*Fault pattern of Cardinal district*

these synclines and of the Crooked Creek anticline is complicated by crumpling and faulting.

The conditions described are not confined to the Cardinal district. Southeast of Elk Creek[1] both the Brazeau syncline and the zone of sheet thrusting have been traced by the writers almost to the northern end of the Brazeau Range, a distance of about 20 miles. The Brazeau syncline passes between the Brazeau Range and the Bighorn Range, but the relation between the zone of low-angle thrusts and the Brazeau anticline is not clearly understood. The Crooked Creek anticline is approximately on the trend of the Brazeau Range, but the writers cannot state that both belong to one axis of uplift.

Reconnaissance examination by the senior author north of the Pembina failed to produce evidence of any definite northern extension of the low-angle thrusts, although the Brazeau syncline is recognizable northwestward almost to the Athabaska River, some 40 miles, and a zone of disturbance exposing the marine Upper Cretaceous rocks along the northeastern flank of the Brazeau syncline may mark the extension of one or more of the thrust faults recognized on the Pembina and southward.

## THRUST FAULTS

### GENERAL DESCRIPTION

The thrust faults of this area and the sheets bounded by them are exposed on the anticlinorium. The relation between them is very clearly displayed along the streams; in the interstream areas enough data are available to permit the drawing of tentative geological boundaries. The probable positions of the traces of the faults underlying the several sheets are indicated in Figure 3.

At least seven of these sheets exist and possibly nine or more since three repetitions of strata at one end of the area cannot be conclusively identified with a similar series at the other end and other faults may exist beneath the lowest series exposed. The thrusts are numbered from the top down as this is probably the order in which they were developed. Each sheet is numbered to correspond with the fault underlying it.

The observed faults make angles with the strata which range from 0° to 20°, the hanging walls generally displaying smaller angles of incidence than the footwalls. Bedding-plane slippage has generally reduced the angle of incidence, locally even to the extreme of causing bedding plane and fault plane to coincide. Drag on the fault planes has tended to increase the angle of incidence, beds in the zone of

[1] Name used by local residents, later called Canyon Creek by Canadian Geology Survey.

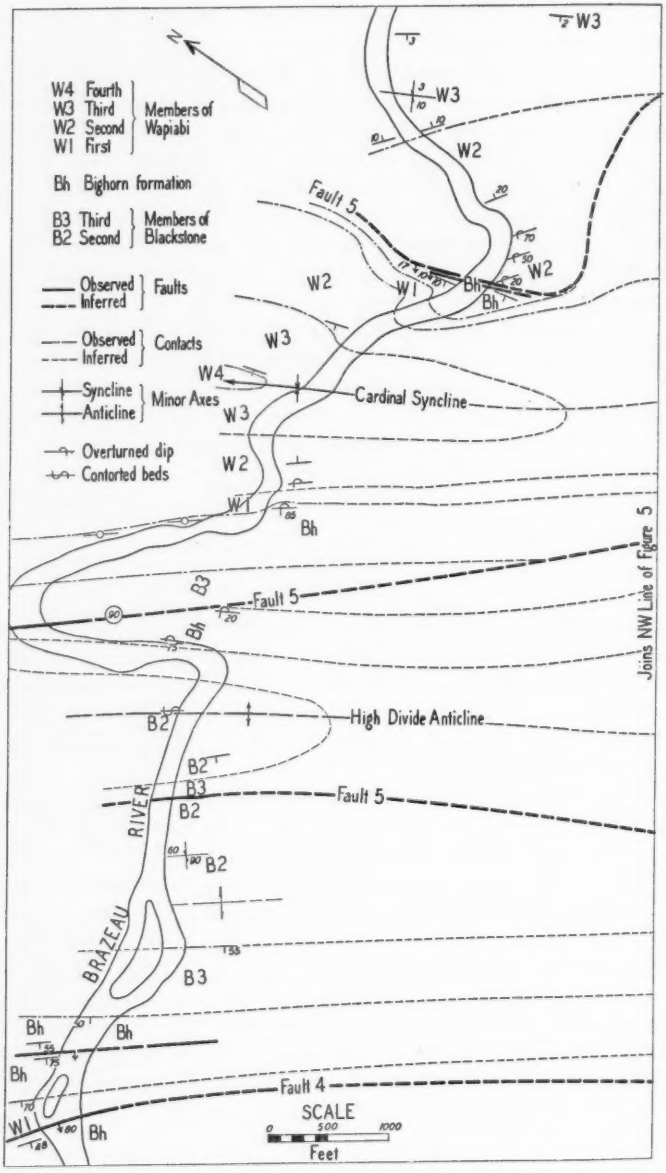


FIGURE 4.—*Areal geology along part of Brazeau River*

drag being locally inverted. Probably the original angles between the faults and the strata averaged about 15°.

Nowhere were these faults observed truncating major folds nor cutting downward through the column in the direction of movement. Without exception they cut upward through the sediments in a northeasterly direction, which is that of the fault movement, and to some extent also in a northwesterly direction along the regional trend of structural axes.

The total shortening of the section by the seven recognized thrusts can be satisfactorily estimated only on one section (Slice Creek and East Pembina, Fig. 10). Here the observed stratigraphic displacements and angles between fault planes and bedding enabled the writers to calculate the total displacement to be about 30,000 feet and the total stratigraphic displacement to be about 7000 feet.

The stratigraphic displacement accomplished by fault 5 is greater on the east flank of the Cardinal syncline than on the western flank of the same syncline and still greater on the eastern flank of the Crooked Creek anticline. Since the angles between the faults and bedding planes are small in all cases the actual displacements along the fault planes must vary in a manner somewhat corresponding with the stratigraphic displacements. The manner in which these variations are produced by bedding-plane slippage and crumpling is discussed in a later section of this paper.

In general the faults are folded in substantially the same degree as the sediments and it is therefore apparent that the displacements took place while the strata were approximate planes. Although it seems certain that any considerable folding of a fault must prevent further displacement along its plane, minor irregularities developed by folding might be sheared off during a subsequent fault movement. Phenomena interpreted as evidence of such shearing have been observed and are described on subsequent pages.

### EVIDENCE OF FOLDED THRUST FAULTS

*General considerations.*—Before the discovery of indisputable evidence of folded thrust faults in this area some of the observed faults were interpreted as upthrusts from the east and some as upthrusts from the west. During that period of the investigation the field relations seemed extremely puzzling and impossible to fit into a reasonable structural hypothesis. When it became apparent that the fault planes had been intensely folded the field data fitted into a systematic whole. Although the structural interpretation presented here is to a great extent the product of inference, the authors are confident that it is essentially correct because it has repeatedly survived that acid field

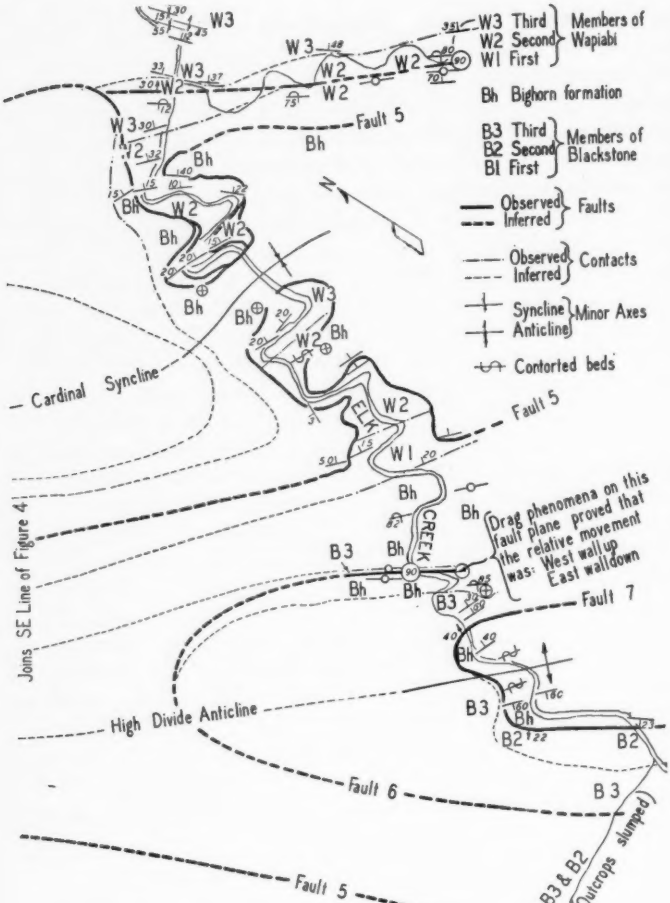


FIGURE 5.—*Areal geology along part of Elk Creek*

test of geologic understanding, the prediction of what is to be encountered on the other side of the hill.

Some of the accompanying figures show the detailed observations in areas where the relations between faulting and folding were most obvious and convincing. The structure sections are composed of a number of short segments, each an accurate representation of observations but joined together to form diagrammatic entities for which purpose the separate segments are shown as though they had been slid into juxtaposition along the strike of the beds (Figs. 6, 10).

*Elk Creek section.*—The lowest of the thrusts (Fig. 5, number 7) is excellently exposed in the canyon of Elk Creek, where it is folded over the High Divide anticline. This fault has been traced continuously between its west dipping and its east dipping exposures, and throughout that distance Bighorn is overlain by Blackstone shale which normally underlies it.

The fault plane was observed dipping 22° SW. on the west flank of the anticline in the canyon of Elk Creek, whereas on the east flank it dips 40° NE. Here then is a complete exposure of a part of a thrust fault which makes a moderate angle with the faulted beds and which has been strongly folded subsequent to faulting. The beds in the hanging wall on the west flank of the anticline are older than beds which occupy a similar position on the east flank, showing that this fault cuts upward through the section toward the east.

Fault 6 (Fig. 5) is likewise well exposed on the east flank of the High Divide anticline, where its attitude is vertical and it causes a repetition of most of the Bighorn. It was at first suspected of being a high-angle thrust from the east, but close examination of the actual contact by the senior author showed that the older beds in the east wall impinged on the fault in a downward direction and were dragged backward and upward against the plane, demonstrating that the east wall moved downward (or eastward if the fault plane were horizontal) relative to the west wall. If fault 6 is folded over the anticline as fault 7 is seen to be, and if the layer of sediments between them is a thin sheet, the root of fault 6 should be present not far west of that of fault 7. A zone of disturbance combined with repetition of the middle member of the Blackstone formation indicates that such is the case, and that this fault also cuts upward in the section from west to east.

East of the High Divide anticline, for about half a mile along Elk Creek, nearly flat-lying strata of Bighorn sandstones are faulted (fault 5) upon east-dipping Wapiabi shales. The fault plane is here folded into

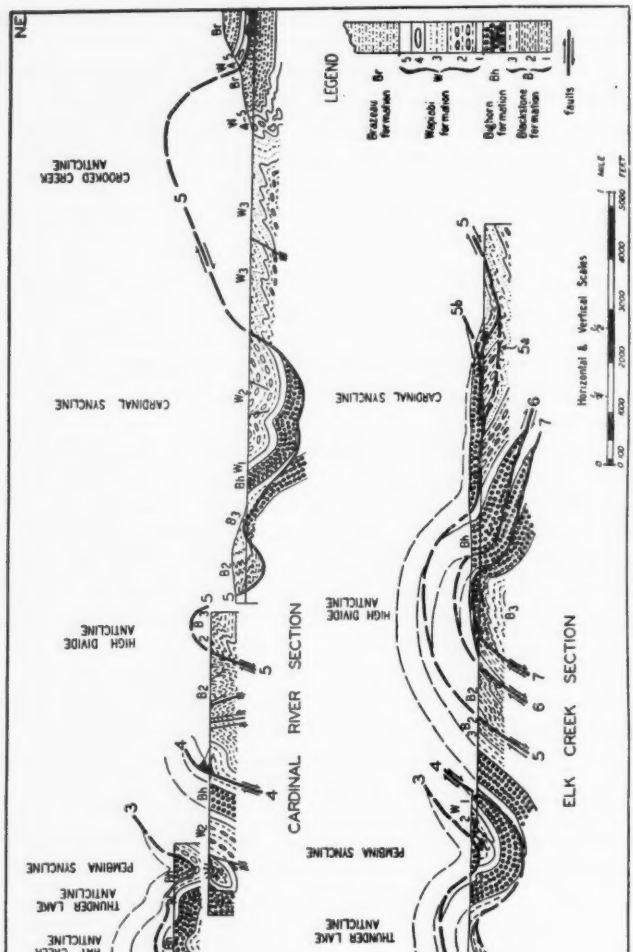


FIGURE 6.—*Geologic sections along Elk Creek and Cardinal River*
Southern part of Cardinal district.

a gentle syncline and can be identified almost continuously in both walls of the canyon for about 3000 feet. The root of this fault is recognized in a duplication of part of the Blackstone formation on the west flank of the High Divide anticline.

Some of the more westerly outcrops on the west fork of Elk Creek (Fig. 2) exhibit relations which can only be explained as the results of low-angle thrusting and subsequent folding. A syncline developed in Wapiabi formation is overlain by a synclinal remnant of Bighorn formation, apparently an outlier of a sheet (number 3), the root of which is concealed on Elk Creek, but exposed on the Brazeau River about 2 miles to the northwest.

The synclinal outlier of fault 5 which is so clearly visible in the canyon of Elk Creek (Fig. 5) was recognized in the canyon of the Brazeau River (Fig. 4). The Cardinal syncline is here plunging northwestward so that this outlier is very wide on the Brazeau and the plane of fault 5 is below the river level under the Cardinal syncline. The root of fault 5 is visible also on the Brazeau River a little farther west, on the west flank of the High Divide anticline.

The migration of this fault upward through the stratigraphic column in the rocks both above and below it, is well displayed in the Brazeau River exposures (Fig. 4). In the most southwesterly exposures this fault places middle Blackstone on upper Blackstone. In its next exposure to the northeast it places upper Blackstone on Bighorn and in its third exposure, still farther eastward, it places Bighorn on Wapiabi.

The roots of faults 3 and 4 were recognized on the Brazeau River as shown in Figure 2, but Figure 4 does not extend far enough southwest to show fault 3. No traces of their planes were discovered farther eastward along that river. The planes of faults 6 and 7, which are exposed in the canyon of Elk Creek, have apparently been carried below the level of the Brazeau River by the northward plunge of the structure.

The observations along Elk Creek and the inferred structure are indicated in the Elk Creek section (Fig. 6). Elk Creek canyon is a narrow gorge with freshly eroded walls 50 to 100 feet high on one or both sides throughout most of its length. The exposures are therefore so nearly continuous that the element of inference in the construction of this Elk Creek section is negligible. The writers are convinced with respect to this section that no materially different interpretation of the observed facts could reasonably be made.

The complex relations in the northeastern end of the Elk Creek section (Fig. 6),—fault 5 branching into 5, 5a and 5b—are discussed and illustrated in a later section.

*Cardinal River section.*—In the western part of the Cardinal River section (Fig. 6), a fault correlated with fault 3 of the Elk Creek section because of its position and the continuity of ridges to the south, is folded into a sharp syncline (the Pembina syncline) in which lowermost Bighorn rests on Wapiabi. Here the beds of the two formations are quite similar lithologically and the strata above and below the fault are parallel to each other but not to the fault surface. The fault itself is not a prominent feature of the exposure, but the abnormal sequence of strata is certain, and upstream the Bighorn is repeated east and west of the root of the fault. The folds here plunge to the north, so that in the cross section it has been necessary to show the fault below river level, where it is folded into the Pembina syncline, although it was actually observed in the river bank.

In the second segment of the Cardinal River section in Figure 6, just east of the Pembina syncline, is a fault which the writers have not been able to recognize on the other streams, and which therefore has not been numbered though it is of the magnitude of the folded thrusts of the district. It may be a branch of either fault 3 or fault 4 and be folded with them in the High Divide anticline. As it outcrops between beds above the Bighorn it probably will be recognized only from outcrops on the streams.

The exposures of fault number 5, from the High Divide anticline east along the Cardinal, indicate that this fault is folded into three anticlines and three synclines. It cuts upward in the section both in the foot wall and the hanging wall from the middle Blackstone shale to the Brazeau formation; its displacement increases eastward from 3000 feet on the base of the Bighorn to 5000 or 6000 feet on the base of the Brazeau, and the direction of movement may be inferred from the position of drag folds in the Blackstone and Wapiabi formations.

The root of this fault was not observed, but its position could be inferred from the extreme crumpling of the middle Blackstone on the west side of the High Divide anticline in a section where the axes of the drag folds incline to the east and indicate a steep west dip. Just to the east the axes of the drag folds incline to the west at low angles. The root of fault 5 is thought to separate these two exposures.

In these outcrops the position of the drag folds with respect to the axis of High Divide anticline is not in accord with the usual relation of such folds to larger ones. Their orientations clearly indicate development by a uniform drag from west to east,—presumably by movement on a low-angle fault—and subsequent modification of positions by folding. Because the Cardinal River runs along the east flank of the High Divide anticline at right angles to the sections it has been necessary

to offset the section at this point in order to avoid showing some confusing relations.

Figure 7 was constructed to display the actual observations in this vicinity on a larger scale because generalizations and omissions were necessary in Figure 2. The outcrops of Bighorn formation near the Cardinal River in the western part of Figure 7, cannot be assumed to continue southeastward because of the almost continuous exposures of other rocks in that direction. The exposures of Bighorn formation in the eastern bend of the River (Fig. 7), are clearly overlain by Blackstone shales brought into contact with the Bighorn by a low angle fault. These exposures of Bighorn formation cannot be connected with the exposures crossing the Cardinal River farther east, which in this vicinity are portions of a continuous ridge of east dipping beds occupying part of the western flank of the Pembina syncline. Undoubtedly the exposure of Bighorn in the easternmost bend of the River (Fig. 7) is due to a window, exposing a folded thrust fault. The correlation of this fault with that just south of the easternmost of three islands (Fig. 7) and both as parts of fault 5 is the writers interpretation, arrived at after hours of consultation and numerous re-examinations of the locality; they are therefore confident that this interpretation is correct. The relations visible in the south bank of Cardinal River at the locality of this window are displayed in Figure 8.

Recumbent drag folds at the right mark the east limb of the High Divide anticline. They straighten up in the syncline, and then flatten again to the east. Fault 5 rises a few feet above river level to reveal distinctive lower Bighorn under the Blackstone. The Bighorn beds diverge from the fault eastward, and drag folds in them indicate that the overriding mass moved in that direction.

The west flank of the Crooked Creek anticline (Fig. 6) discloses a steep fault which places lower Wapiabi (W 2) over strongly crumpled middle Wapiabi (W 3). In the east flank of this anticline a low-angle thrust plane dips gently to the east and places upper Wapiabi on Brazeau beds. The beds above this plane are not crumpled, whereas those below are thrown into large drag folds, the axes of which incline to the west and prove the eastward movement of the overlying mass.

Figure 9 shows the field relations of these exposures which, although scattered over a considerable area, are nevertheless sufficient to prove the existence of the low-angle fault and its identity north and south of the River, as indicated on Figures 2 and 3 where the Cardinal River approaches the margin of the illustration.

The correlation of this low-angle thrust with the steep thrust and identification of them with the various parts of fault 5 to the west,

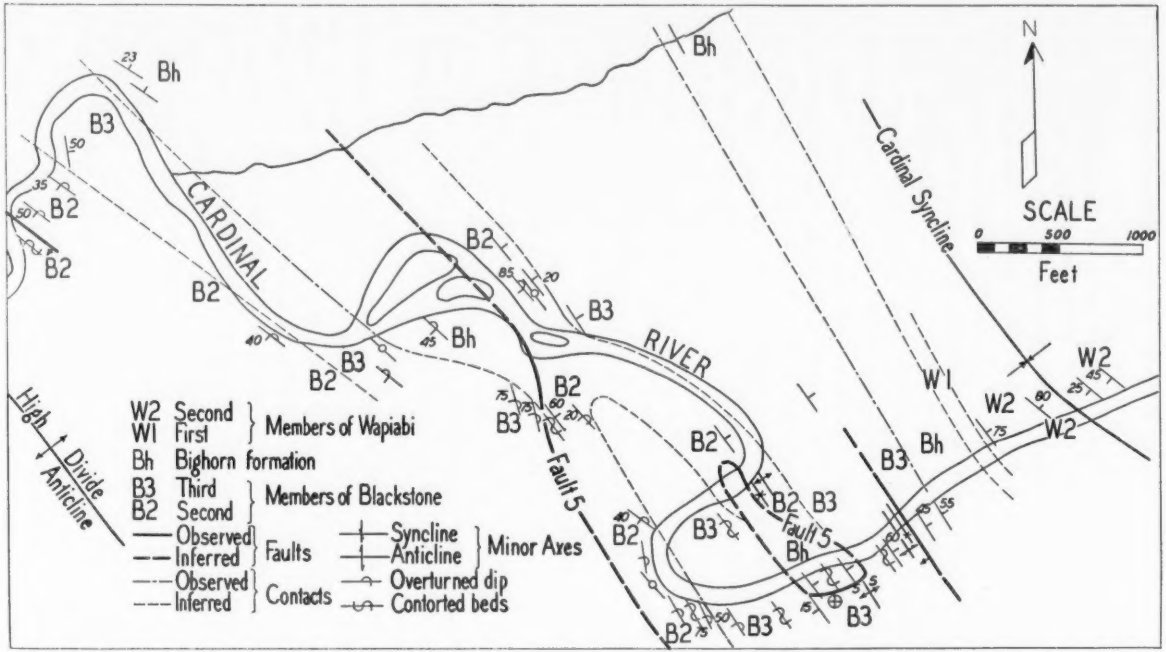


FIGURE 7.—*Areal geology along Cardinal River near High Divide*

(*see* Fig. 6) is the logical explanation. The crumpling of the beds in the footwall of this fault where it cuts the Bighorn and Wapiabi formations is apparently the reason why its displacement is less on the western limb of the Crooked Creek anticline than on the eastern limb of that fold, where the beds in the walls of the fault are relatively little deformed. This cross section (neglecting the difference in scale) resembles sections

FIGURE 8.—*Sketch of an outcrop on Cardinal River*
Showing relations between folded thrust fault and drag folds in faulted Cretaceous sediments. Approximate length of section: 800 feet.

through the Calcareous Alps, (Heim, 1919) showing systematic drag folding in weak sediments under comparatively undeformed nappes.

In the Cardinal area drag folding is generally confined to certain members of the formations, notably the sandy, thin-bedded middle zone (number 2) of the Blackstone and the platy shale zone (number 3) of the Wapiabi, both of which are susceptible to crumpling on account of their composition of alternating thin layers of stronger and weaker rocks.

*Slice Creek and East Pembina section.*—On the east flank of the Rat Creek anticline the Bighorn is repeated three times, in each case overturned to the east (Fig. 10). The eastern occurrence forms a syncline above which, in a ridge, is a fourth outcrop of Bighorn essentially vertical. These relations were inexplicable until the general nature of the thrusting was understood but now it can be easily accounted for. The continuity of the Bighorn in the syncline indicates the existence of a nearly flat fault (designated as number 1) separating the Bighorn in the syncline from the vertical Bighorn beds higher in the ridge.

Figure 11 shows the actual field observations along Slice[2] Creek in this vicinity. The upper (more easterly) of the two faults which repeat the overturned Bighorn is designated as number 2 and the lower is correlated with fault 3 of the Cardinal River section.

The roots of these faults were not observed, but a great excess of west-dipping Blackstone section on the west flank of the anticline in-

[2]Now called Rat Creek by the Topographical Survey of Canada, although when this work was done that name was locally applied to another stream. (*See* Figures 2 and 3 for stream names used herein.)

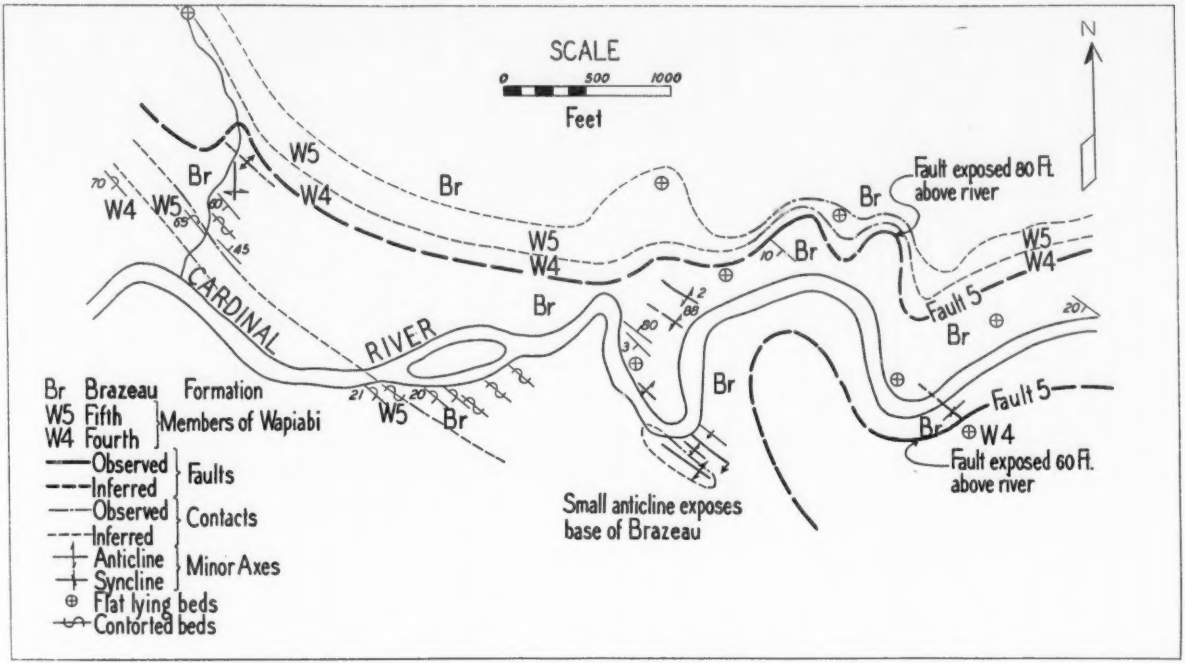


FIGURE 9.—*Areal geology near mouth of Cardinal River*

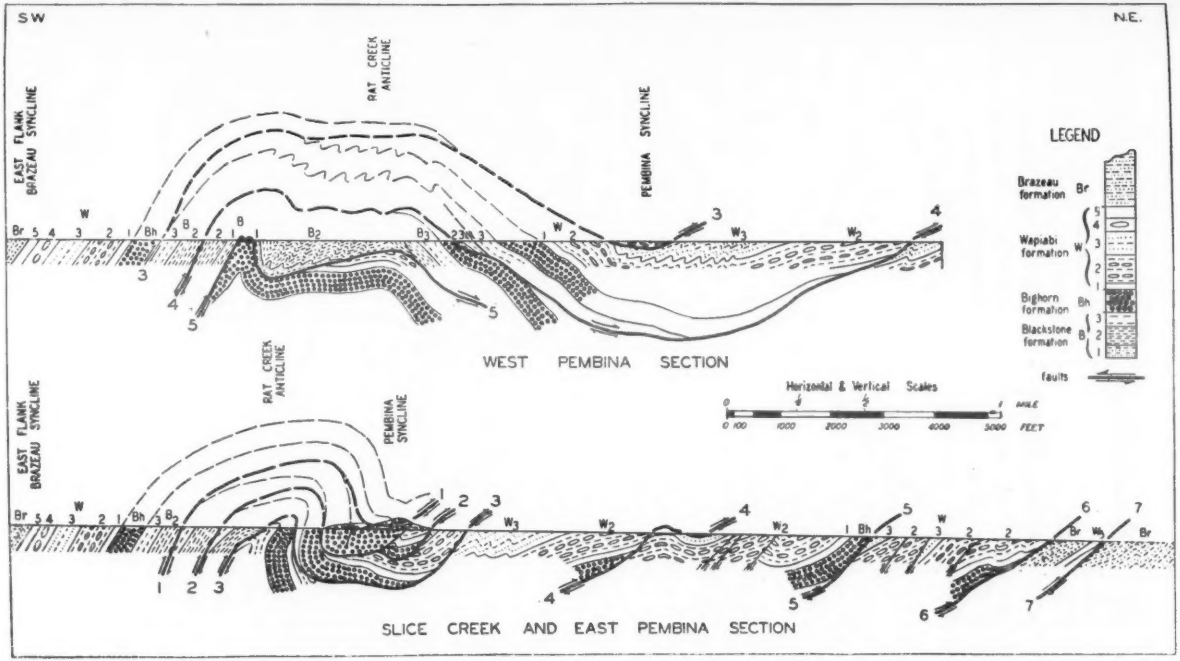


FIGURE 10.—*Geologic sections along Slice Creek and Pembina River*
Northern part of Cardinal district.

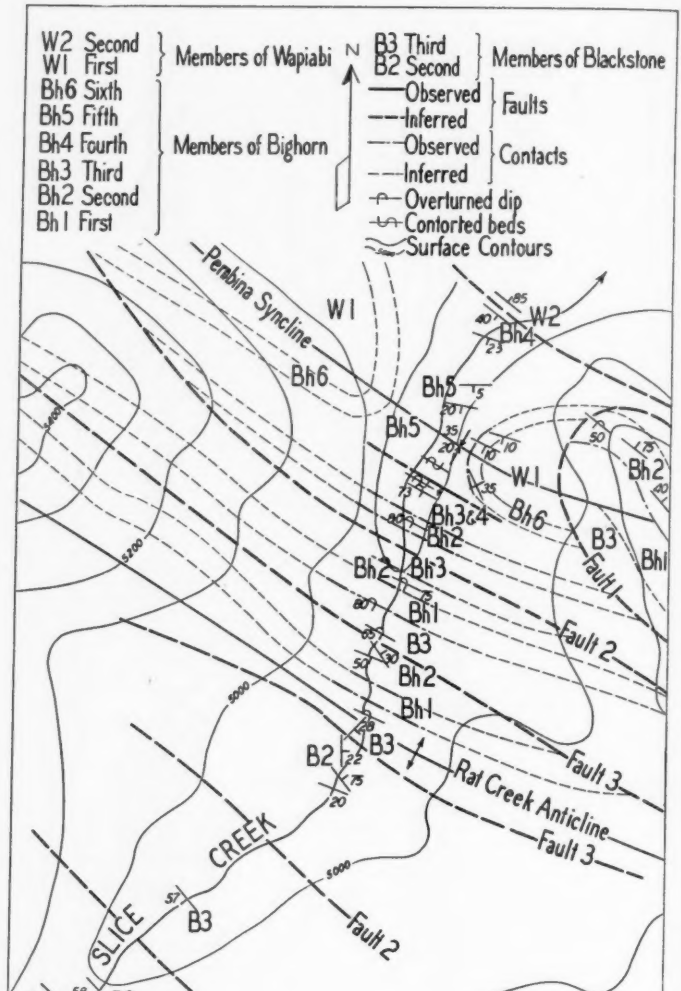


FIGURE 11.—*Areal geology along Slice Creek near axis of Pembina syncline*

dicates the position of at least two and probably three high-angle thrusts. Correlation of these with those which repeat the Bighorn is in accord with the observed systematic relations between the faults and the strata in other parts of the area.

Tentative identification of faults 2 and 3 on the east flank of the Pembina syncline was made by projection of the beds downward from the observations, in accordance with the structural habit of the thrusts as revealed in other sections. This procedure yielded a result in harmony with field relations where these faults should emerge but were not actually exposed.

A series of closely adjacent outcrops on the Pembina River and Slice Creek near their confluence displays four fault contacts which are interpreted as exposures of fault 4 where it has been folded into a flat anticline and a shallow syncline. Figure 12 shows the field relations between the several exposures in this locality.

The structural interpretation is shown in the Slice Creek and east Pembina section (Fig. 10). The fact that in each one of these closely adjacent fault exposures the second zone of the Wapiabi formation was thrust over the third zone and the fact that on each one of these exposures the material in the footwall of the fault was more crumpled than was the material in the hanging wall seem, to the writers, strong evidence that these exposures are segments of one folded thrust.

*West Pembina section.*—On the axis of a small anticline in middle Blackstone beds are two exposures of Bighorn sandstone (Fig. 10). The outcrops are small, and might pass unnoticed, as the beds are strongly crumpled, but they point definitely to the presence of fault 5. The sharp anticlinal buckle which causes this window (Fig. 14) in sheet 5 marks the beginning of a narrow faulted uplift which extends northward for more than 25 miles beyond the limits of the Cardinal area.

A thrust recognized on the Pembina River and correlated with fault 4 of the other sections, is indicated by a duplication of part of the Blackstone shale west of the window referred to above. The extension of this fault under the Pembina syncline is shown by a duplication of the Bighorn on the west flank of that structure, (Figs. 10, 13, 14) and by the reappearance of the fault in the Wapiabi formation on the east flank.

The northern tip of the large outlier of sheet 3 is recognized on the axis of the Pembina syncline. In Figure 13 field observations along part of the Pembina River indicate several exposures of a fault correlated with fault 3. At this point the fault is folded into the Pembina

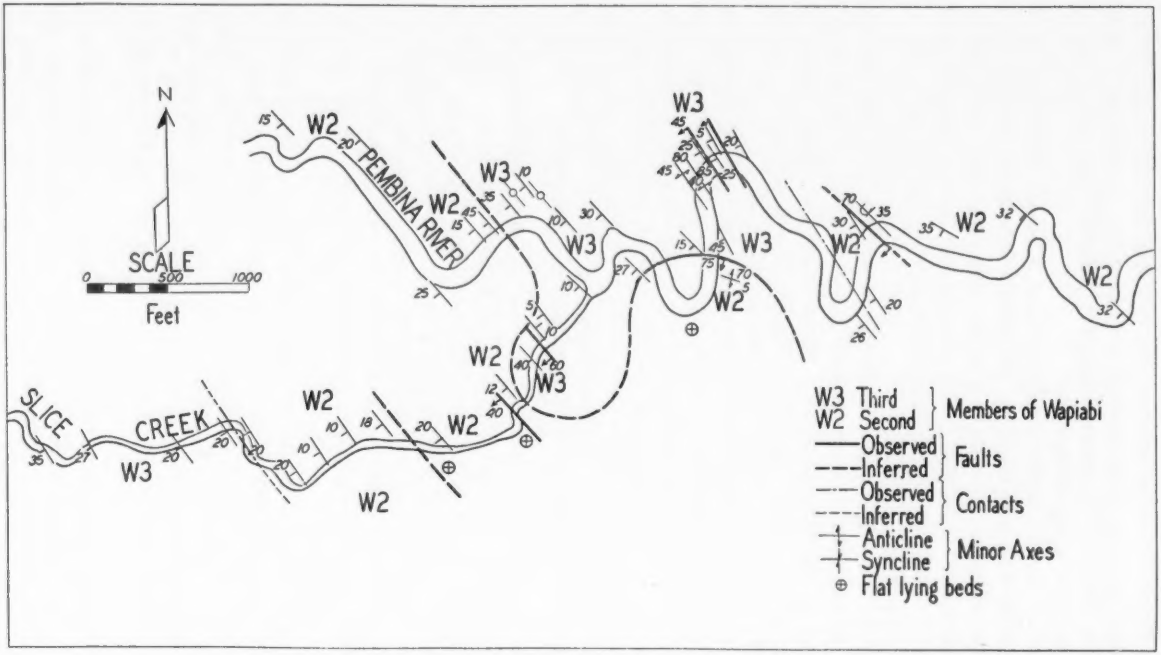


FIGURE 12.—*Areal geology near confluence of Pembina River and Slice Creek*

syncline which is here plunging southeastward. The several closely adjacent exposures of the fault and the very definite identification of the different parts of the Wapiabi formation which occurred in its hanging wall and footwall left no doubt in the minds of the writers concerning the correctness of the interpretation which is shown in the cross sections in Figures 10 and 13.

The beds beneath this fault are here strongly crumpled while those above it are much less disturbed; therefore, the writers infer that its displacement is here increasing eastward. This is one occurrence of a synclinal remnant of a thrust sheet for which no root is shown in Figure 10. This fault must either merge with fault 4 over the Rat Creek anticline, or pass westward into a bedding plane as indicated in Figure 10. The latter is more probable and is entirely in accord with the theory of development presented in this paper.

No traces of faults 1 and 2 are found on the Pembina River. In part this is due to the southward plunge of the folds in this vicinity, but the disappearance of the roots of these faults is thought to result from a merging of their planes and a diminution of displacement on them, together with local replacement of faulting by bedding-plane slippage.

*Hills North of Pembina River.*—In the hills north of the Pembina River the Bighorn sandstone is repeated in three shallow synclines which lie like dishes one above the other (Fig. 14). The lowest Bighorn "dish" overlies outcrops of the Brazeau formation. On the west flank the two upper "dishes" can be traced down into the exposures of Bighorn on either side of fault 4 on the River. On the east flank the exposures run out, presumably against the underlying fault planes as these rise in the section to the eastward. These Bighorn "dishes" furnish perhaps the most striking evidence of the extent and thinness of the individual thrust sheets.

## FEATURES RELATED TO FOLDING

The folds developed in this faulted mass of sediments are fairly steep and locally closely compressed. Minor folds add to the complexity of structure in some localities and on account of the offset relations produced by the major thrusts the surface axes of the principal folds are not continuous. A notable amount of rather high-angle minor thrust faulting, especially in synclines, is believed to have been produced by the folding, subsequent to the development of the major thrusts. Folding has likewise altered the dips of the axial planes of the drag folds developed during the earlier major thrusting and has probably augmented and in some places originated crumpling.

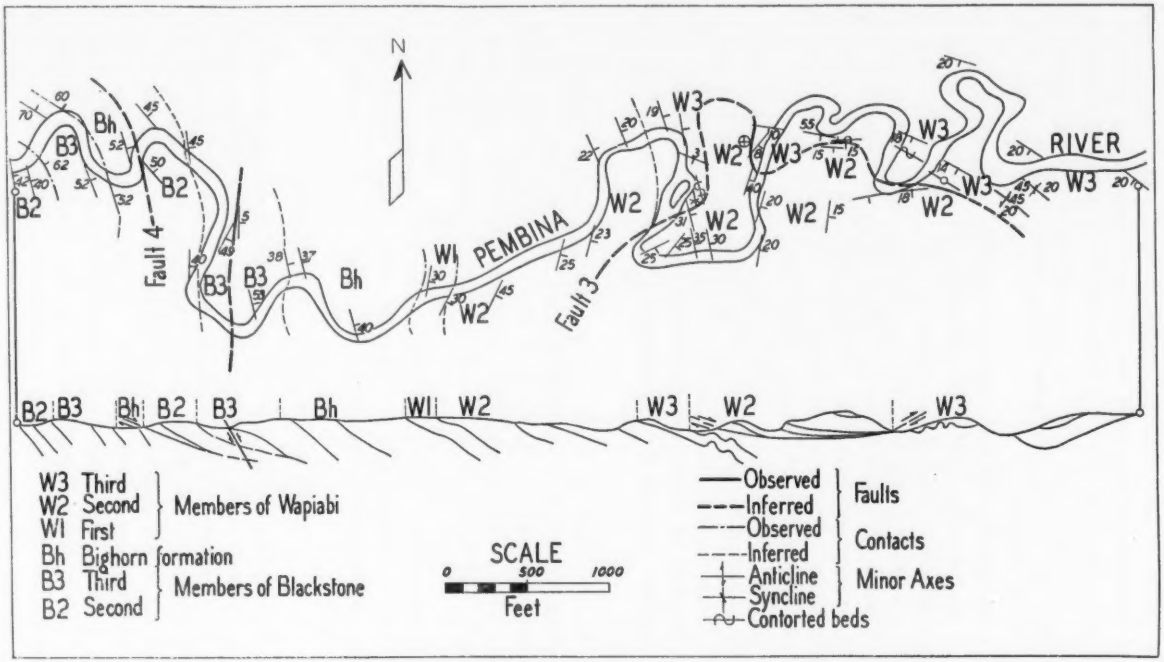


FIGURE 13.—*Areal geology along Pembina River near axis of Pembina syncline*

### DRAG-SHEARING ALONG FAULT PLANES

On lower Elk Creek beneath thrust 5, parts of four fault blocks or slices are exposed (Fig. 15). The drag on the minor faults and the

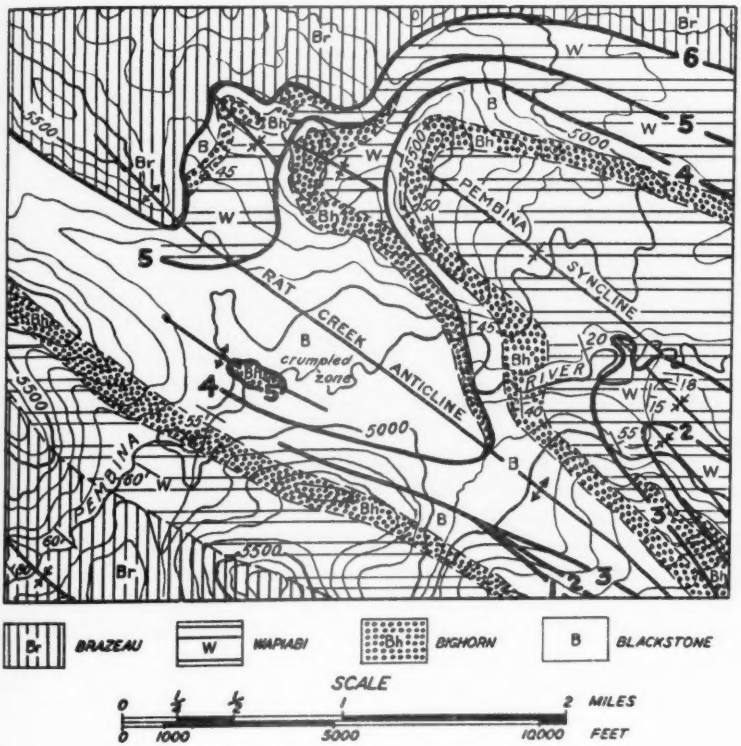


FIGURE 14.—*Sketch map of northern end of Cardinal district*
Showing topography and details of areal geology.

continuity of fault 5 above them show that the former are not thrust from the east.

The writers believe that faulting and folding in this area were somewhat overlapping events, so that the thrust planes were slightly folded between the later increments of fault movement. Where a thrust plane was bowed upward (Fig. 15, diagrams *a* and *b*) subsequent movement sheared off a portion of the footwall and carried it forward. Where a thrust plane was bowed downward (diagrams *c* and *d*) sub-

sequent movement sheared a slice from the hanging wall and the slice was left behind. Several combinations of folding and subsequent shearing through footwalls and hanging walls have been tested, but the only one which explains all of the observed facts is that shown in Figure 15.

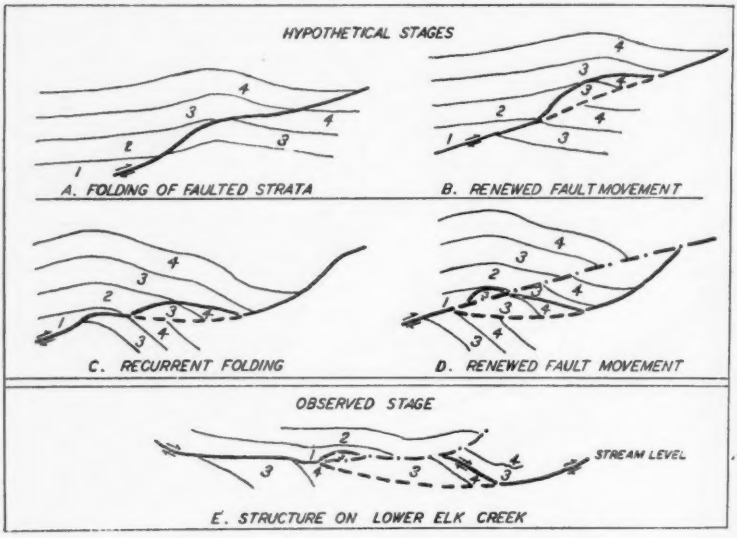


FIGURE 15.—*Diagram to explain development of fault slices*
Observed on lower Elk Creek.

### DRAG FOLDS

The writers believe that drag folds developed during the formation of the thrusts, that the east limbs of such drag anticlines are generally the shorter, and that their axial planes bear a systematic relation to the planes of the thrust faults which controlled their development. Originally the axial planes inclined uniformly to the west, and where they are now recumbent, vertical or dipping to the east, their position has been modified by later folding of the thrust sheets. The great majority of observed drag folds are of this origin but in some cases, as in the asymmetrical syncline on the west flank of the Rat Creek anticline (Fig. 10, Pembina section) they may have been accentuated by or even originated during the later folding.

### MINOR THRUST FAULTS

Numerous minor faults are shown on the cross sections (Figs. 6, 10) and nearly all of them are overthrust up dip, away from the axes

of the synclines. Those formed in east-dipping beds, therefore, show motion opposed to that of the larger faults. They are thought to result from a process similar to that which formed the thrust sheets, acting on a much smaller scale during the development of the local folds. In no case are they thought to be imbrications of the sheets as is postulated for Highland structure.

## REGIONAL HABIT OF FOOTHILL STRUCTURES

### PREVIOUS DESCRIPTION

Thrust faulting is an outstanding feature of the geologic structure of the Rocky Mountains and their foothills in Alberta. Most writers appear either to have regarded folding and faulting as independent phenomena or to have thought of folding as subsidiary to faulting.

Dowling (1914) published a preliminary description of the structure in the vicinity of the Turner Valley anticline, near Calgary, showing a number of thrust faults, though his examination was not detailed and he did not suggest the complexity of structure which drilling has revealed.

Structure sections in many of the published works on this region show that the authors believed that many of the complicated folds and faults which they observed could not reasonably be supposed to continue to great depths but must die out downward in some fashion. In his description of the Moose Mountain District, Cairnes (1914) suggested in his structure section that the visible beds were cut by faults which did not dislocate some of the more deeply buried strata.

Stewart (1919) traced a number of folds and faults through an extensive area in the foothills of southwestern Alberta, but his structure sections show the faults to be essentially vertical and with uniform displacements at different depths, a conception opposed by the published conclusions of later workers.

Slipper (1921, p. 5) and Hume (1926) in describing their observations in and near the Turner Valley oil field further emphasized the role played by thrust faulting. MacKay (1930, p. 16) examined the Cardinal area during his study of the coal fields of the surrounding region, but with respect to the Cardinal district the writers cannot entirely agree with his structural interpretation.

In his paper on the structure of Turner Valley, Goodman (1930) emphasized the idea that the formations nearer the surface were faulted and folded discordantly with the structure of the more deeply buried rocks. His structure sections also displayed his belief in the tendency to intense crumpling and development of recumbent folds in incompetent rocks in synclines.

Hume (1930, p. 15) advanced the conception that the character of deformation undergone by sediments in this region was to some extent dependent upon the nature of the sediments themselves. In the structure sections which accompany a more general discussion of thrust faulting in Alberta, Hume (1931a) showed graphically that displacements on these thrust faults vary with depth, but he did not discuss the phenomenon. In a later publication, Hume (1931b, p. 52B) developed the idea that the total displacement on foothill faults may be many times the stratigraphic displacement and that thrust faults in this region may disappear into bedding planes.

One of Link's (1931) cross sections shows a fault passing downward into a bedding plane in Paleozoic limestone. It is obvious, however, from a subsequent cross section (Link, 1931, p. 505) that he does not consider such fault to remain or to die out in the bedding plane, as he shows all faults in this latter cross section becoming steeper at greater depths and maintaining their approximate parallelism.

That it is difficult to construct reasonable sections of Canadian foothill structures without showing the displacement on the thrust faults as varying in magnitude at different levels is shown by one of MacKay's (1933) sections in which his "bed No. 8" is nearly parallel to the plane of the fault identified by the "Mutz outcrop" whereas lower strata are truncated at steeper angles. The fault reaching the surface between Hillcrest and Bellview is drawn as a thrust fault with a considerable displacement near the surface of the ground and as a fault with exactly the opposite displacement in almost the same magnitude near the base of his section.

## GENERAL STRUCTURAL CHARACTERISTICS

Individually or together during about 8 years the writers studied the geologic structure throughout the foothills of Alberta and from this experience have formulated a conception of what may be termed the regional structural habit. A brief statement of the ideas involved is a desirable preface to the discussion of the origin of the low-angle thrusts described herein.

If the strata of the faulted anticlinal ranges and synclinal valleys of the Rocky Mountains and their foothills in Alberta were restored to a condition of low dip, they would constitute a series of major folds with wave lengths across the strike of from 5 to 15 miles. However, the anticlines are relatively narrow, compressed, asymmetrical, and usually faulted, whereas the synclines are broader and less compressed.

The major uplifts are sometimes anticlinoria. The Brazeau Range is a group of related anticlines. On the head of the north fork of the Berland River (T. 53, R. 6, W. of the Sixth Meridian) the senior

author observed a limestone range composed of two closely appressed anticlines in heavy bedded Paleozoic limestones, each fold nearly closed and with a brecciated core.

Apparently, during the development of these structures there was a persistent tendency for the anticlines to be more intensely compressed than the adjacent synclines, and to develop asymmetrically with a steeper eastern limb. Probably each major anticline increased in size until it reached the limit imposed on its growth by the strength of the heavy-bedded Paleozoic limestones within it and then failed by shearing of its steeper flank. That many of the thrust faults so developed now stand at high angles near the surface is attested by the rather straight traces of such faults across the rugged topography which they frequently traverse.

Subsidiary faulting and folding of strata in both anticline and syncline accompanied the major deformations. The zone immediately in front of a large thrust is often seen to be intensely crumpled and faulted. Such is the case in front of the Bighorn, Brazeau, and Nikanassin ranges. The gentler limbs of asymmetrical structures are commonly cut by thrust faults which dip in the same direction as the strata they dislocate. An excellent example of such a fault cutting into the Paleozoic limestones and passing downward into a bedding plane was observed by the senior author near the head of the south fork of Cabin Creek on the west flank of the Cabin Creek Range, near the southwestern corner of T. 55, R. 4 W. of the Sixth Meridian. Such faults are believed to result from bending of the limestone strata under lateral compression which increased in intensity upward. They are, therefore, considered genetically similar to the thrusts described in this paper.

### COMPETENCE OF PALEOZOIC SEDIMENTS

The massive Paleozoic limestone series is believed to have dominated the development of visible structures, but this series seems to have had a lower limit of competence at or near the base of the Devonian sediments, for although many ranges expose Devonian rocks, rarely are any pre-Devonian rocks brought to view. Moreover, the anticlines are not broad arches like those of the Rocky Mountains in Montana, Wyoming, and Colorado which are known to contain granite cores; they are narrow, compressed structures like those in southwestern Wyoming and southeastern Idaho. The strata within them dip so steeply and the anticlines are so narrow that it seems to the writers impossible to consider them as surface manifestations of folds involving any great thickness of the earth's crust. The probable confinement of this type of folding in this region to a thin layer of the earth's crust was postulated by

Hume (1931b), but he appears to believe that folding of sediments in the foothills of the Canadian Rockies is subsequent and incidental to the development of large thrust faults, a conclusion which the writers cannot wholly accept.[3]

The writers believe that at or near the base of the Devonian there was a regional zone of parting or *décollement* which permitted the Devonian and younger sediments to deform differently from the older rocks; that folding was the process by which the post-Silurian section received its initial deformation; and that faulting generally took place when and where the sediments could no longer accommodate themselves to the applied stresses by folding alone.

### INCOMPETENCE OF MESOZOIC ROCKS

During the formation of the major anticlines and subsequently as these faulted and overrode the adjacent synclines, the latter were considerably reduced in width. The resulting intense squeezing of the Mesozoic sediments is regarded as one of the features which dominated the development of structure in those rocks.

The second factor of primary importance in their structural development is the relative incompetence of these sediments. The general character of structures developed in them shows clearly that except where supported by the Paleozoic limestones the Mesozoic rocks older than the Brazeau formation have been unable to form an arch more than a few thousand feet wide. Consequently as the major synclines were compressed the Mesozoic strata in them buckled and sheared in ways which were only indirectly related to the structure of the underlying limestones.

### IMPORTANCE OF STRONG BEDS AND BEDDING-PLANE SLIPPAGE IN MESOZOIC COLUMN

In spite of the general incompetence of at least the lower portion of the Mesozoic column, it contains some strata of massive sandstone which could within limits act as struts and transmit stress for some thousands of feet. Such beds occur in the Brazeau and Bighorn formations and also in some of the Lower Cretaceous formations.

These more rigid members, when under excessive loads, yielded to compression by shearing and thus initiated fault planes at angles which permitted relief of stress. Where the superincumbent load was not too

[3] Since this manuscript was written Hume (1939, 1940) has published papers on the structure of areas in the foothills of Alberta in some of which he shows faults which have sinuous traces on his cross sections, and thus suggest the possibility that the fault planes have been folded, but his texts do not indicate that he attaches much importance to such a possibility. However, he does state definitely (1940, p. 6) that a fault is known to pass into a bedding plane and thus cease to cause stratigraphic displacement, and that "the steep prevalent dips of the strata in the foothills does not truly represent conditions at depth within the more competent Paleozoic limestone." Link and Moore (1934) also recognized the fact that fault planes have been subjected to "warping or folding."

great they yielded by competent folding, though on a smaller scale than that of the primary competent folds originating in the Paleozoic beds. The stronger of the Mesozoic strata thereby localized and controlled the minor anticlines and synclines. Wells drilled for oil and located on surface anticlines, which penetrated unexpected faults, or failed to find the subsurface formations reflecting the surface structure, bear out this idea. Thus folding in the competent Paleozoic rocks was coincidental with faulting in the lower part of the Mesozoic section just above them, and with folding on a smaller scale in upper Mesozoic rocks nearer the surface. A given series may have faulted and crumpled incompetently while heavily loaded and then folded after erosion had sufficiently reduced the load. The strong folding of the faulted sediments described in this paper is a case in point, though this sequence of events may have been due to a change in the nature of the stress conditions rather than to unloading.

The folding of any stratified series requires a great amount of bedding-plane slippage, or its equivalent in crumpling or shearing, and this requirement was of primary importance in the development of structural types in the foothills of Alberta and especially important in the development of thrust faults in the Cardinal district.

## DEVELOPMENT OF LOW-ANGLE THRUSTS

### PRIMARY FOLDS AND ORIGIN OF THRUSTS

The writers believe that these thrusts are the product of the type of folding which prevails in the region and are not offshoots from nor the surface expression of any deep-seated major thrust fault. The repetitions of the Bighorn formation by the observed thrusts involve an upward movement of at least 7000 feet and a northeastward movement of some 30,000 feet. The search for the origin of such a displacement leads one inevitably to the uplift of the Nikanassin and Bighorn ranges, because only this uplift seems adequate.

Most outlier and spur ranges seen by the writers in the foothills of Alberta are closely compressed, more or less overturned anticlines with limestone cores. At a given stratigraphic horizon the apex of such a fold may be from 15,000 to 25,000 feet above the adjacent synclines. The width of such an anticline is of the same order as its height, and a thrust fault of moderate displacement is a characteristic feature of its eastern flank.

The Nikanassin Range is this type of fold and the writers know no reason to suppose that this fold rose through its cover of Mesozoic rocks on a thrust plane. On the contrary it probably developed as an asymmetrical arch, moving northeastward and lifting competent strata into

its steeper flank until some late stage of its development. This "rolling up" of the strata is regarded as the cause of successive closely spaced thrust faults.

If the folded strata of the Cardinal district and the foothills westward to the Nikanassin Range are graphically restored to low dip angles, the major thrusts may then be represented as planes. On the assumption that these thrusts do not cut the Paleozoic rocks, it appears that the shortening of the section by the thrusts (30,000 feet in the Bighorn formation) furnishes us with an excess length of limestone sufficient to draw a deep overturned syncline at the eastern base of the Nikanassin Range. The syncline thus constructed embraces the apparent roots of the thrusts and is of proper proportions to be in conformity with the regional character of the limestone folds; the assumption of the existence of such an overturned syncline does not conflict with what is known of the geology of the eastern flank of the Nikanassin Range.

### ASYMMETRY AND CONCENTRATION OF STRESS

If all strata involved in the upfolding of the Nikanassin anticline had been competent to lift superincumbent loads, the folding there developed would have been of the similar type in which thickening of beds over the anticlinal axis and thinning on the flanks permit equivalent amounts of shortening of section in all stratigraphic horizons. Since the Mesozoic rocks were incompetent to lift the superincumbent load a fold of the concentric type developed in which the section was shortened less across the anticline in higher strata than in beds nearer to the competent Paleozoic strata. The rise of an anticline of this type produced in the syncline on its steeper flank an intense compressive stress which increased upward and had to find relief through bedding-plane slippage, crumpling, or shearing.

Figure 16 illustrates the amount of bedding-plane slippage required of beds in the gentler flank of an asymmetrical syncline in a section which has been bent through 90 degrees. The total slippage required in a section of a thickness similar to that in the Cardinal district is of the order of magnitude of the displacement on the observed overthrusts.

Presumably the slippage (or equivalent deformation) was concentrated in the gentle limb of the syncline because the beds in the asymmetrical Nikanassin anticline became so sharply bent at an early stage in the growth of the fold that slippage of strata westward up the steeper flank of the syncline and across the axis of the anticline was physically impossible. Thus all the material having to be squeezed away from the axis of the syncline was of necessity forced up the gentler limb of that syncline.

As indicated in diagram *b* of Figure 17, the amount of material which must be so squeezed up the gentler flank from the interior of a sharply folded syncline is a relatively large mass. Where the beds in such a syncline are incompetent, this transfer of sediments cannot take place by bedding-plane slippage but must be accomplished by crumpling or

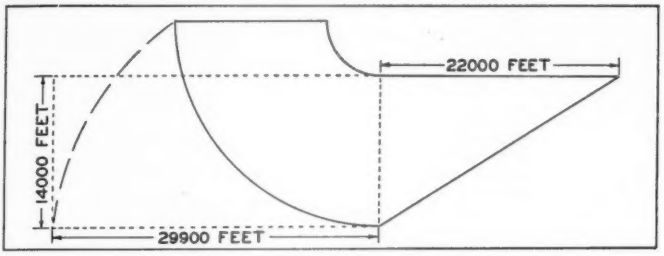


FIGURE 16.—*Bedding-plane slippage*

Possibly required by folding of a thick stratified series. To accommodate a fold of 90° in a block of sediments 14,000 feet thick if all bedding-plane slippage is concentrated in one limb, it is necessary for uppermost bed to slide forward 22,000 feet relative to lowermost bed.

shearing or both. In the Cardinal district the transfer was accomplished by a combination of crumpling and thrust faulting within the limits placed on each process by the nature of the deforming sediments, and by the conditions surrounding the deformation.

### DEVELOPMENT OF THRUST FAULTS

The development of this type of shearing and crumpling is thus a function of the rotation of the limbs of the folds; the maximum transfer of material being out of the syncline and away from the limb which suffered the greater rotation. In the Mesozoic section within the curve of the syncline east of the asymmetrical Nikanassin anticline a powerful compressive stress tended to squeeze material out of the syncline. The strong sandstone members in the section prevented the Mesozoic sediments from yielding entirely by crumpling or plastic deformation. Each of these strong members acted as a strut transmitting the stress eastward along the bedding until the resistance to such movement exceeded the crushing strength of the stratum, which then sheared. Compressive stress parallel to the bedding within these strong members, modified by shearing stress set up throughout the deforming mass by the rotation of the steeper west limb, produced low-angle thrust faults cutting the sediments in the gentler eastern limb of the syncline.

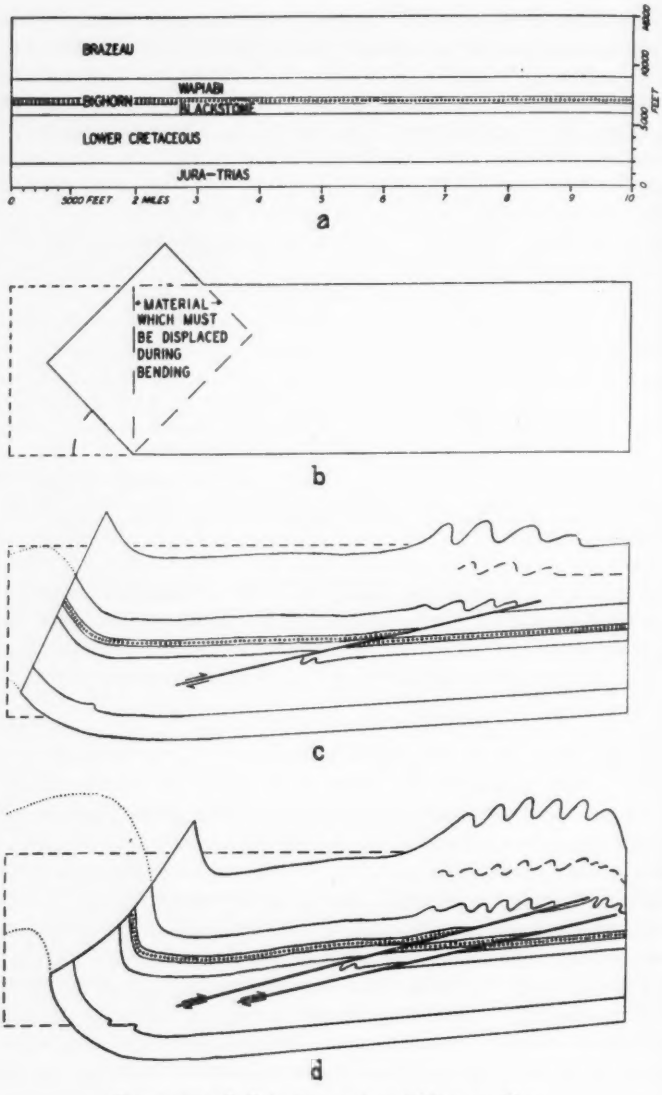


FIGURE 17.—*Relief of stress in a folding syncline*

By development of low-angle thrusts on the gentler flank, showing manner in which displacement on such faults may decrease both in direction of movement and in the opposite direction.

Diagrams *c* and *d* of Figure 17 were made to illustrate the writers' notion of the progress of this type of faulting. In the construction of these diagrams the lengths of all strata were kept constant and the areas of the sections are constant and equal to that of the undeformed section (diagram *a*) so that the bending of the strata and the displacements on the faults have quantitative significance.

Displacement on such a fault increases upward (dies out downward) where bedding-plane slippage extends the hanging wall or where crumpling shortens the footwall. It decreases upward (increases downward) by the reverse of these processes. Figure 16 shows that if it were not for the mass of strata to the east, bedding-plane slippage could entirely relieve the shortening in the syncline. The low-angle fault plane frees the ends of the overlying strata and permits relief in the manner illustrated in Figure 17. The fault is thought to originate in the resistant strut, and to develop upward and downward, with displacement diminishing away from the locus of maximum displacement, which may be also the locus of initial rupture.

In diagrams *c* and *d* of Figure 17, the increase in displacement from the root of a fault to the Bighorn formation is largely though not entirely accomplished by bedding-plane slippage of the hanging wall, whereas the decrease in displacement above the Bighorn is accomplished by crumpling of the hanging wall. In the Cardinal district crumpling was observed both above and below the faults, with more below than above. It appeared to be largely confined to certain laminated stratigraphic units which were especially susceptible to it. Displacement on these faults, therefore, may locally increase and decrease according to the characters of the faulted beds.

Where the angle between fault plane and bedding plane is small, displacement on the fault may be transferred to a bedding plane. Crumpling of the underlying strata may permit development of a large actual displacement without stratigraphic offset. The sudden disappearance of faults 1, 2, and 3 at the north end of the area seems to require this explanation.

### REASONS FOR SUCCESSIVE THRUSTS

One thrust fault might suffice to relieve the stress in the axis of a folding syncline and doubtless such has been the case in many localities. However, in the Cardinal district at least seven such thrusts were formed, and to account for this fact the writers again refer to the "rolling up" of the steeper flank of the syncline.

During the growth of the syncline east of the Nikanassin Range the locus of maximum curvature in the competent bed was migratory. As the steeper limb grew in height this locus migrated eastward. The mi-

gration advanced the position of the plane of shear in the gentler flank of the syncline; existing faults no longer sufficed for the relief of the stress and a new fault was formed.

The fact that at least seven such thrusts involving a total displacement of 30,000 feet formed in a zone about 5000 or 6000 feet in thickness shows the great necessity for shortening with a relatively small amount of lateral migration of the locus of maximum curvature. This condition would arise from the formation of an overturned syncline with a short radius of curvature.

### INCREASE OF DISPLACEMENT IN OLDER FAULTS

Probably the movement on earlier faults did not cease with the development of subsequent ones, for the bedding-plane slippage requirements increased throughout the overthrust mass with each increment of folding of the major syncline. Therefore, the hanging wall of each fault must have been progressively elongated and displacement correspondingly increased, unless this requirement was satisfied by crumpling. The necessity for additional displacement on an early fault after the development of a subsequent one was forcefully brought to the writers' attention during the construction of the constant-area diagrams in Figure 17.

### SPACING OF THRUST FAULTS

The spacing of the shear planes of the low-angle thrusts is thought to be a function of the elasticity of the various sandstone struts in the Mesozoic series. Had the material all been shale the eastward migration of the locus of bending and of the plane of shear would have produced a series of closely spaced shear planes with insignificant displacements on each. Had the material been more rigid throughout, failure would probably have occurred by folding rather than by low-angle thrusts.

## INITIATION OF FOLDING

Since the faults are folded in almost the same degree as the sediments, the latter must have been but little disturbed when faulted. Some change in the mechanism of the major deformation appears necessary to change the method of relief of stress. The attitude and character of the folds suggest that during folding compressive stress was more uniformly distributed vertically through the section, that the rotational stress was less effective, and that the marine Cretaceous strata were relatively more competent than during the thrusting. These conditions would be produced by the growth of a major anticline in the competent beds below or east of the zone of thrusting, and by erosional unloading occasioned

by the thrusting and by the growth of this anticline. Evidence of the beginning of such a fold is found in the field data.

The Brazeau formation in the Brazeau syncline above the highest of the thrust faults in this area is only slightly elevated above rocks of the same age situated on the downthrown side of the lowest of these faults. Therefore, apparently the sediments of the Cardinal district have been generally tilted about 10° SW., or prior to the low-angle thrusting, a regional dip of this magnitude and direction existed in the Brazeau formation. The writers favor the former alternative, and for this reason some downwarping of the major syncline is indicated in diagrams *c* and *d* of Figure 17. The accentuation of the anticline of which this dip is the west limb was the event which caused the cessation of thrusting and the inauguration of folding.

The writers suggest that the faulting of the eastern limb of the Nikanassin and Bighorn anticlines diminished and perhaps halted the "rolling up" of the competent strata in the steeper flank of the major syncline and thus not only caused the reduction or cessation of the stresses which produced the low-angle thrusting, but also transferred regional compression more completely to deeper horizons and favored development of competent folds farther eastward. Thus diminution of faulting and increase of folding in the Cardinal district may have been events more or less synchronous with the faulting of the Nikanassin Range.

## CONCLUSION

### RECAPITULATION OF THEORY

The foregoing argument may be summarized as follows:

(1) The mechanism here postulated depends on the existence of an incompetent series above a competent series.

(2) In such a sequence the incompetent beds fold concentrically over the anticlines and therefore are shortened there less than the competent series.

(3) This requires excessive shortening of the incompetent beds in the syncline and this requirement increases upward throughout the incompetent series.

(4) In an asymmetrical syncline this shortening of the incompetent beds is concentrated on the gentler limb.

(5) The stress developed there consists of compression parallel to the beds combined with rotational stress. These produce low-angle shearing under appropriate conditions of load.

(6) The growth of a competent asymmetrical anticline causes forward migration of the synclinal axis, which advances the zone of critical shear-

ing stress and develops successive closely spaced low-angle thrust faults.

(7) The rigidity of the stronger members in the incompetent series controls the transmission of stress and the localization of shear, and thus determines the spacing of the thrust faults.

(8) As there is less demand for shortening of the competent series than of the incompetent strata in the synclines, and as the competent series is by far the stronger, it seems probable that such faults die out above the competent beds.

(9) As superincumbent load diminishes upward in any section, the upper portion of the so-called incompetent series may have been competent to relieve the stress by folding rather than by shear. The low-angle thrusts therefore may never have reached the surface until exposed by erosion.

(10) Variation of displacement, necessitated by (8) and (9) is brought about by bedding-plane slippage which may cause either lengthening or shortening of a wall of a fault and by crumpling which causes a shortening of the crumpled wall. The amount of displacement developed on any fault is locally diminished by crumpling of either or both walls. The displacement is locally increased where bedding-plane slippage decreases the angle between the bedding planes and the fault plane. Conversely the displacement is locally decreased where bedding-plane slippage increases the angle between bedding planes and fault plane.

(11) Elevation of the mass by deformation promotes erosional unloading above the zone of thrusting, which lowers the zone of folding in the incompetent series. Forward progress of major folding develops a new competent anticline which counteracts the rotational stress. These factors combine to end low-angle thrusting and start the development of folds involving the fault planes as well as the faulted beds. The shearing of lenticular slices from the walls of the low-angle faults shows that folding started before fault movements had entirely ceased.

(12) During the development of this folding stress of similar origin to that which produced the low-angle faulting was developed in the synclines, and caused minor shear thrusts which formed dipping in both directions toward the axes of the synclines.

(13) The shortening shown by the folding of the competent series on the major anticline should not be added to that shown by the zone of low-angle faulting, to give the total reduction of width of the deformed belt. The two are equivalent manifestations of the same movement in different parts of the column. Furthermore total shortening of the deformed zone is only equivalent to the sum of fault displacement and related crumpling when the two are measured at the same stratigraphic horizon.

(14) The authors suggest that the mode of origin of this differential or rotational stress may find an application in the explanation of sheet thrusting, even where it is developed on a much larger scale.

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TERTIARY ROCKS
(BIDAHOCHI FORMATION)

Diatremes filled with pyroclast

Volcanic necks (basalt-filled)

Basalt interbedded with Bid

Sedimentary rocks of Bidaho

GEOLOGIC
(Bas

Dikes

pyroclastic rocks

-filled)

with Bidahochi formation

f Bidahochi formation

MESOZOIC ROCKS

Mesaverde sandstone (Upper Cretaceous)

Mancos shale (Upper Cretaceous)

Dakota sandstone (Upper Cretaceous)

Triassic and Jurassic rocks (Chinle, Navajo, and Morrison formations)

LOGIC MAP OF MAJOR PORTION OF HOPI BUTTES AREA

(Based on photoengravings of U. S. Soil Conservation Service)

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# SEDIMENTATION AND VOLCANISM IN THE HOPI BUTTES, ARIZONA

BY JOHN T. HACK

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## ABSTRACT

The Hopi Buttes area in the southern Navajo Country, Arizona, lies in a Pliocene basin of sedimentation and volcanism. The thin Pliocene sediments, known as the Bidahochi formation, are interbedded calcareous sand, clay, marl, tuff and other pyroclastic material. These sediments extend southeastward as far as the Zuni area, New Mexico. They overlie the Hopi Buttes erosion surface of low relief, a portion of which is now exhumed on the southern edge of Black Mesa. This surface may be widespread in northern Arizona. The Bidahochi formation and the erosion surface have evidently been deformed in Pliocene or post-Pliocene time.

The Hopi Buttes area, which lies at the lowest point of the structural basin of deformation, contains Pliocene volcanic rocks of alkalic composition greater in volume than the sediments, and over 200 closely spaced volcanic necks or diatremes ranging from 500 to 4000 feet in diameter, and arranged along a complex pattern of fractures.

Flows are few in number, for the volcanism produced pyroclastic débris and lava domes. Deep erosion reveals that explosions produced funnel-shaped pipes which

were rapidly filled with explosion débris, with material poured in by streams or with lava and viscous agglomerate erupted from below. Small rims protected many diatremes for a time from filling by river-borne sediments so that fine ash, gypsum, and calcium carbonate collected in their crater lakes to be later buried by coarser material as stream aggradation continued. The structure of the diatreme fillings indicates that subsidence occurred after eruption.

FIGURE 1.—*Index map*

A—area of reconnaissance studies, shown in Figure 2.
B—Hopi Buttes area, shown in Plate 1.

## INTRODUCTION

The portion of the Navajo Country known as the Hopi Buttes is dominated by eroded and dissected lava-capped buttes, and volcanic necks filled with basalt and pyroclastic debris. The volcanic necks, or diatremes, are over 200 in number and closely spaced.

Several factors make this area favorable for observation of the structure and origin of diatremes. The volcanism in the Hopi Buttes occurred in the Pliocene when the area was receiving a thin cover of sediments

derived from highlands on the north and east. These sediments and the interbedded volcanic debris overlie an erosion surface of low to moderate relief, known as the Hopi Buttes surface. At the present time the Pliocene rocks and the erosion surface under them are dissected in places to depths of over 700 feet, so that structures of typical diatremes can be observed at the level of their uppermost extremities as well as at depths far below the original level of extrusion.

The Hopi Buttes surface thus serves as a reference plane from which to measure the depth of erosion of the various diatremes. This erosion surface and the sediments which overlie it are widespread in northeastern Arizona. Evidence of deformation of these sediments and the underlying surface is revealed, which is important both for correlation and for an understanding of the nature of the volcanism.

## PREVIOUS WORK

Geologic work in the Hopi Buttes has been limited to reconnaissance studies. Gregory (1917) describes the general geology and petrography in his comprehensive work on the Navajo Country. Williams (1936) has published an account of the Hopi Buttes volcanic field in a report on the petrography of the volcanic rocks of the whole Navajo Country. Reagan (1932) named the Pliocene sediments of the Hopi Buttes the Bidahochi formation. The most important features of the region have thus been briefly treated, but the ideal conditions for the study of volcanic structures and geomorphic history have not been fully interpreted and discussed.

## ACKNOWLEDGMENTS

The present study began while the writer acted as geologist for the Awatovi Expedition of the Peabody Museum of Harvard University. Mr. William H. Claflin IIIrd, a member of the expedition staff, ably assisted the writer in the field for several weeks. Kirk Bryan, E. S. Larsen, Jr., and M. P. Billings visited the writer in the field and offered many valuable suggestions. The field work was completed under a generous grant from the Penrose Bequest of the Geological Society of America and greatly aided by the generous hospitality of Mr. and Mrs. Clarence N. Halderman of Indian Wells, and Mr. and Mrs. Wilmer C. Roberts of Jeddito Trading Post.

The mapping was based on photoengravings of aerial mozaics and base maps furnished by the U. S. Soil Conservation Service.

## TOPOGRAPHY

The Hopi Buttes form a highland mass rising above the barren plains of the southern Navajo Country to an average altitude between 5500

and 6500 feet. The average relief is only about 500 to 700 feet, but their rise from the flat, alluvium-filled valleys gives the buttes an august appearance. This region may be divided into: (1) The high, central area which includes the rolling surface of the Black Rock Buttes,—an area of unexcavated diatremes, lava domes, and small lava flows which give a false appearance of a continuous but irregular lava-capped plateau; and (2) the surrounding lower area which is more deeply dissected and consists of scattered volcanic necks or high lava domes perched on rocky buttes.

Extensive highlands lie to the north and east of the volcanic area. Black Mesa, a dissected plateau 100 miles in diameter, overlooks the Hopi Buttes from the north. On the east a less conspicuous highland underlain by the Pliocene Bidahochi formation stretches eastward to the Defiance Plateau.

## PRE-TERTIARY ROCKS

The Pliocene lavas and sediments rest unconformably on Mesozoic sedimentary rocks which include the Cretaceous Mesaverde sandstone, Mancos shale, and Dakota sandstone; the Jurassic Morrison formation and Navajo sandstone; and the Triassic Chinle shale. These rocks dip slightly to the north so that the oldest rocks crop out in the southern portion of the area. In general the region rises to the north in a series of broad but low benches or escarpments where each resistant bed crops out, culminating in the plateau of Black Mesa. This more or less irregular ascent is broken by the mass of volcanic rocks which forms a highland in the Hopi Buttes. The major portion of the Hopi Buttes area is shown in Plate 1.

## PLIOCENE BIDAHOCHI FORMATION

### GENERAL DESCRIPTION

The volcanic rocks are interbedded with water-laid marl, calcareous tuff, calcareous clay, and calcareous sand which on Roberts Mesa are about 400 feet thick. These sediments have been named the Bidahochi formation by Reagan (1932); Wilmarth (1938) accepted this name. According to Williams (1936) and Wood (1941) the formation is Pliocene.

### AREAL EXTENT

The Bidahochi formation is apparently thickest on Roberts Mesa. It extends a short distance to the north in small outliers on the flat surface of Black Mesa (Fig. 2), and extends southward into the Hopi Buttes where it is made up mostly of pyroclastic debris interbedded with lava

domes. Sedimentary material is absent in the Buttes at the southwestern margin of the area. The formation extends continuously eastward from the Hopi Buttes beyond Ganado where it overlaps the westward dipping

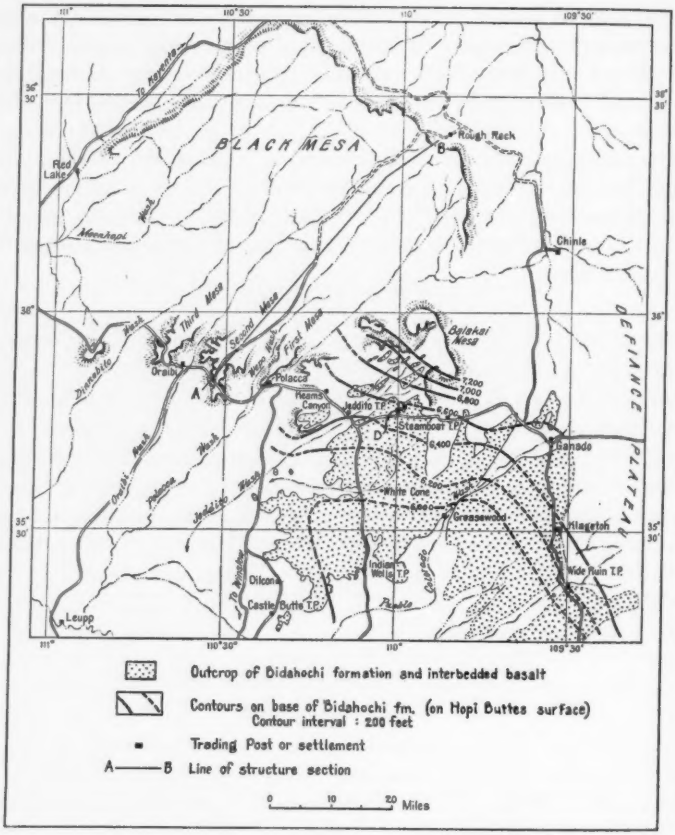


FIGURE 2.—*Map of part of western Navajo Country*
Showing outcrop of Bidahochi formation and Hopi Buttes surface. Based on map of Navajo Country of Office of Indian Affairs.

Permian rocks of the Defiance Plateau and is found southeastward beyond the limits of the Navajo Country. The Bidahochi is apparently continuous with the late Tertiary rocks of the Zuni area of New Mexico, described by McCann (1939).

LITHOLOGY

*General statement.*—Volcanic rocks comprise the major portion of the Bidahochi formation among the volcanoes of the Buttes. To the north and east, however, sediments predominate, although fine-grained tuffs are apparently universally present. The lithology varies considerably depending on the source of the material.

*Sediments.*—On Black Mesa the Bidahochi formation is confined to small outliers less than 100 feet thick. It is composed chiefly of calcareous sand with thin concretionary layers of coarse, crystalline calcium carbonate. The most abundant sand grains of this facies are subangular quartz resembling closely the sand of the Mesaverde sandstone. Plagioclase and grains of microgranular quartz, both minor constituents of the Mesaverde sandstone, occur also. However, even where the proportion of sand is over 80 per cent, the rock looks like a sandy limestone. It is almost certain that in the northern area of outcrop on Black Mesa and on Roberts Mesa the Bidahochi formation is composed largely of erosion products from the Mesaverde sandstone.

South of Black Mesa, on Roberts Mesa and near White Cone, the Bidahochi formation is about 350 to 400 feet thick. The white calcareous sand constitutes only the upper 100 to 200 feet of the formation. It overlies much finer-grained pink and green calcareous clay, as shown in the following section:

SECTION OF BIDAHOCHI FORMATION ON ROBERTS MESA

North of White Cone

(Elevation of top of section is 6530 feet)

| | Feet |
|---|---|
| (1) Thin layer of white, chalky marl or limestone caps Roberts Mesa | 2 |
| (2) White calcareous sand with some clay or bentonite and containing hard concretionary layers | 64 |
| (3) Interbedded greenish-gray and pink sandy clay | 12 |
| (4) White calcareous sand with numerous calcareous concretionary layers | 53 |
| (5) Interbedded greenish-gray and pink or red sandy clay | 24 |
| (6) White calcareous sand | 11 |
| (7) Interbedded greenish-gray and pink sandy clay | 24 |
| (8) Greenish-pink calcareous sand with concretionary layers | 10 |
| (9) Dark greenish-gray, coarse basaltic tuff or sand, well stratified | 3 |
| (10) Interbedded greenish-gray and red sandy clay, calcareous | 64 |
| (11) Covered | 10 |
| (12) White, rhyolite tuff, quarried by natives | 3 |
| Total | 280 |

Base of formation is covered.

On Bidahochi Butte, the type locality, the calcareous sand is entirely absent, and the lower beds of the formation are composed of brown, greenish-gray, or reddish, calcareous clay, in places containing gypsum

crystals (Pl. 1). This facies is overlain by coarse pyroclastic rocks or lava flows.

SECTION OF BIDAHOCHI FORMATION

On the north side of Bidahochi Butte

| | Feet |
|---|---|
| (1) Dense monchiquite with phenocrysts of olivine | 50 |
| (2) Pinkish-gray, tuffaceous sandy clay, interbedded with grayish-green, tuffaceous clay which weathers to a gray-white powder | 196 |
| (3) Brown, sandy clay containing abundant rounded pebbles of gray and yellow quartz, especially at base | 34 |
| Total | 280 |

At the southern margin of the Hopi Buttes area the sediments thin, and at Long Butte they are absent. The white calcareous sand present on the south side of Black Mesa is probably the most characteristic facies of the Bidahochi formation, for it is found as far eastward as Ganado and southeastward at Chambers.

At Greasewood the basal beds are apparently similar to the calcareous clay found at White Cone. At Ganado the basal beds are composed of soft brown sandy conglomerate or gravel. At Wide Ruin Trading Post, where the Bidahochi formation rests against the Defiance Plateau, yellowish sand and gravel are dominant.

*Volcanic rocks.*—Volcanic rocks are present almost everywhere in the Bidahochi formation although they are abundant and predominant only near the lava domes and diatremes which are their source.

The volcanic activity was violently explosive, and extrusion of lava was limited to small mushroom-shaped domes around the numerous diatremes. Few flows are continuous. There is little difference, if any, in the composition of the lava of the flows, domes, and diatremes. It is dense, black basalt, in most places limburgite or monchiquite, as described by Williams (1936). The typical monchiquite, probably the most common rock, has a groundmass of microlithic augite and iron ores with interstitial analcite, although some basalts contain interstitial orthoclase, rather than analcite. Phenocrysts are augite, and in some rocks, olivine. In many places, especially in the volcanic vents, the lava is vesicular and the open spaces are filled with zeolites or analcite, which may be replaced by calcite. Chlorite and iddingsite are common secondary minerals.

Pyroclastic rocks are greater in volume than the lavas, and in places make up the entire Bidahochi formation. They range from fine ash to coarse agglomerate containing bombs or blocks of country rock several feet in diameter.

Petrographic study of these rocks has shown that their igneous constituents are similar. The highly variable appearance is due largely to

great differences in texture, varying content of country rock, or to admixture of sedimentary materials. The coarser pyroclastic rocks show bedding only vaguely, or not at all. The finer rocks, however, are well bedded almost everywhere, apparently reworked by streams, and thus are water-laid. The various grades of pyroclastic material can be summarized as follows:

FINE WHITE RHYOLITE TUFF: Occurs over a very large area in one thin bed near the base of the Bidahochi formation,—south of Ganado, at Greasewood, on Roberts Mesa, and in the Bidahochi Valley north of Indian Wells. This tuff was probably derived from a distant source and may be in part water-laid.

CALCAREOUS TUFF: White, thinly laminated, fine-grained ash (in places finer than .005 mm.) altered to calcium carbonate. It occurs as filling in diatremes, probably in part deposited in standing water.

COARSE BASALTIC TUFF: Coarse, sandy greenish-gray tuff. Widespread in Bidahochi formation interbedded with sediments. It is the dominant facies in central area of Hopi Buttes. Almost everywhere this material is well stratified (Pl. 2, fig. 3). Typically it is composed of fragments of basalt glass (limburgite), augite, olivine, and iron ore which range from .10 to 3.0 mm. Fragments of quartz, plagioclase, and microgranular quartz are almost always present, proving the presence of some sedimentary material. In some cases the larger fragments are rounded lapilli, often showing prominent rims. This basaltic tuff is most common as a water-laid pyroclastic sediment interbedded with calcareous clay, but it also fills many diatremes, especially those which are relatively little dissected by erosion.

TUFF-BRECCIA: Coarse, dark-green breccia and basaltic tuff of varying size containing mostly fragments of basalt or minerals of basalt ranging from .1 mm. to several cms. in diameter. This material is nowhere stratified. It fills numerous diatremes.

BASALTIC AGGLOMERATE: Many of the diatremes, especially those now deeply eroded, contain very coarse agglomerate composed of large bombs and irregular masses of vesicular basalt of monchiquitic composition. In many places this agglomerate contains fragments of tuff, clay, and blocks of country rock and is permeated by stringers of dense basalt. It contains crude bedding or what is more probably flow banding.

The basaltic agglomerate was derived by violent explosion which threw out bombs and lapilli or comminuted a solid mass of basalt. This material fell back into the open explosion hole, probably while still hot and viscous, and was later welded by more quiescent intrusion of dense basalt. Often the agglomerate was pushed out in large flat domes beyond the limits of the source vent.

EXPLOSION BRECCIA: Many of the deeply dissected diatremes contain a filling of coarse angular blocks (5 feet in diameter) of Mesozoic rocks mixed with minor amounts of basaltic tuff, tuff-breccia, and sedimentary materials. The most common of the angular blocks are Jurassic sandstone.

Large blocks of Mesozoic and older rocks are also found in the finer pyroclastic rocks near and inside diatremes. In an area near Crazy Waters Spring many large bombs of graphic granite were found, apparently derived from deep-seated pre-Cambrian rocks.

### SOURCE OF MATERIAL AND CONDITIONS OF DEPOSITION

The Bidahochi formation thickens and becomes coarser to the north and northeast. On Black Mesa and immediately to the north it contains material derived from the Mesaverde sandstone, which is present on Black Mesa and which once occurred farther east in the Chinle Valley. Undoubtedly, the sediments of the Bidahochi formation were deposited by streams flowing from the north and east. The pyroclastic material was derived from the diatremes of the Hopi Buttes.

Fresh-water shells occur in the clays of White Cone, a fact which led Williams (1936) to refer to the formation as the Hopi Lake Beds. Ripple-marked and mud-cracked sand and clay do not commonly occur but have been found. The formation contains numerous remains of camel, horse, and mastodon, and a few plants.

During the field work, the writer conferred with Mr. Howard Scott Gentry, a collector employed by Dr. Childs Frick. Among other finds Mr. Gentry showed the writer a skeleton of a small camel which rested on its belly on a thin layer of sand. Its fore feet projected down through the sand into soft greenish clay beneath. This animal had certainly perished because it had broken through a thin layer of sand into wet clay, too soggy to support its weight.

Thus the lithology and biota of the formation suggest deposition by southwestward-flowing streams, on a low, wet, swampy plain. Damming of the streams may have been caused in part by the growth of numerous small volcanoes rising in little hillocks above low marshes and lakes. The bedded nature of the pyroclastic rocks suggests that much of the volcanic material was spread over this low plain by streams in flood. The abundant animal life indicates that vegetation was lush and the climate warm and moist for at least part of the year.

### DEFORMATION OF BIDAHOCHI FORMATION

The base of the Bidahochi formation has been contoured (Figs. 2, 14) with a Paulin altimeter checked two or three times daily when possible at bench marks of the U. S. Coast and Geodetic Survey. The maximum error may in a few places be as great as 50 feet. On the southeastern edge of Black Mesa on Balakai Mesa and Antelope Mesa, the base of the Bidahochi formation slopes to the southwest at an average of 100 feet per mile. In the White Cone area where the base of the formation is composed of clay, it slopes at about 20 feet per mile. Evidently the slope was imposed after deposition, for the material is too fine to have been deposited on the existing gradient.

Deformation is particularly evident between latitudes 35° 30′ and 35° 45′ at the longitude of White Cone (Pl. 1; Fig. 3, section). The Bidahochi formation slopes to the south at a gradient of over 20 feet per mile. At the base it consists of clay beds deposited in ephemeral lakes or swamps. The modern Beshbito Wash which traverses this area from north to south has the same average gradient as the Bidahochi formation, for it crosses the base of the formation at two places, and its alluvium contains coarse gravel, coarser than any material found in the Bidahochi formation in this area.

A thin bed of coarse, water-laid tuff furnished further evidence of deformation. This horizon may be traced all the way from White Cone to the north end of Roberts Mesa (Fig. 3).

It now slopes southward 10 feet per mile. Its source obviously lay to the south, and it must have been deposited on a surface which sloped to the north or was nearly flat. Therefore, the Bidahochi formation in

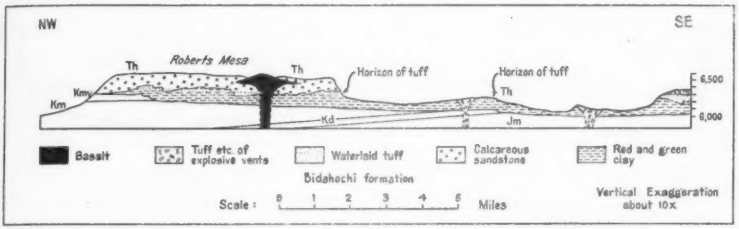


FIGURE 3.—*Structure section X — Y (Pl. 1) across part of Hopi Buttes area*

this region probably has been tilted to the south at least 10 feet per mile, and at least half of the present gradient resulted from Pliocene or post-Pliocene deformation.

## HOPI BUTTES EROSION SURFACE

### GENERAL DESCRIPTION

The Bidahochi formation overlies an erosion surface of low relief which truncates rocks ranging from Triassic to Upper Cretaceous. Gregory (1917, p. 121) recognized this surface calling it the Hopi Buttes peneplain.

In the Hopi Buttes area this erosion surface became the site of sedimentation and intense volcanic activity. At least the southern tip of Black Mesa was also the site of Pliocene sedimentation. The largest part of the mesa, however, rose above the Pliocene lowlands. Apparently only a small part was ever reduced to an erosion surface of low relief. The extent and nature of the erosion surface is a problem of importance to the Tertiary geology of the region, for many volcanic areas in the Plateau region are thought to rest on a supposedly Pliocene erosion surface, or peneplain, and the region about the Hopi Buttes is one of the few having Tertiary fossils.

### HOPI BUTTES SURFACE ON BLACK MESA

*General considerations.*—Black Mesa is a dissected highland about 60 miles in diameter. It is drained by the Tusayan Washes which rise at its northern edge and flow southwestward through steep-walled valleys to the fingering southern escarpment. The north and northeast sides of the mesa

rise to altitudes of over 8000 feet and support stands of scattered yellow pine. A bold and towering escarpment borders Chinle Valley on the east. The southern side has elevations from 6200 to 6500 feet and lies at the lower edge of the zone of juniper and pinyon.

The mesa, viewed from the south and east, presents a wide and even skyline which appears to truncate its structure, and at first inspection

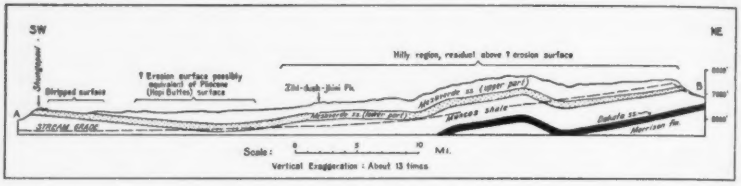


FIGURE 4.—*Structure section A — B (Fig. 2) across Black Mesa*
Elevations determined with Paulin altimeter, controlled from bench mark near Shungopovi and subsidiary barometric control points.

the divides between the southwestward trending canyons appear to be remnants of the Hopi Buttes surface. A traverse across the mesa from Oraibi to Rough Rock (Fig. 2), however, shows that only a small portion of the mesa was ever eroded to a plain of low relief.

*Stratigraphy.*—The soft Mancos shale is exposed on the lower slopes of the escarpment of Black Mesa on all sides. The Mancos is overlain by the Mesaverde sandstone which caps the entire mesa surface, except in the deep interior valleys.

At the southern edge of the mesa, the mesaverde consists of about 50 feet of white, massive sandstone at the top, underlain by about 130 feet of interbedded shale, soft sandstone, and coal, underlain in turn by about 70 feet of buff, massive sandstone. These three divisions, about 250 feet in total thickness, are recognizable throughout the Black Mesa area, and form a recognizable lithologic unit of the Mesaverde, resting on the Mancos shale. To the north the sandstone of this lower part of the Mesaverde increases at the expense of the shale, and at the northern edge of the mesa, the lower part of the Mesaverde is dominantly sandstone.

Except for scattered patches of higher beds, the lower part of the Mesaverde holds up the southern portion of Black Mesa. Northward on Second Mesa, however, (Fig. 4, line A-B) a higher part crops out on the divide between the Oraibi and Wepo Washes. This part is less reduced by erosion northward. North of Zihl-dush-jhini Peak it is over 600 feet thick. This part of the Mesaverde is composed mostly of shale, with thin sandstone and coal beds. The proportion of sandstone in the

upper part increases northward, and at the north end of the mesa the highest beds of the upper part are predominantly massive sandstone, which forms a high and bold escarpment. The contact between the top white sandstone of the lower part of the Mesaverde and the basal shale of the upper part forms a horizon marker which probably can be recognized anywhere on Black Mesa.

*Structure.*—The strata of Black Mesa form a structural basin known as the Tusayan downwarp, whose axis lies near the settlement of Pinyon. At the south end of the mesa (First, Second, Third, and Antelope Mesas) the rocks dip slightly northward and at the north end they dip more steeply southward.

North of Zihl-dush-jhini Peak (Fig. 4) is a flat-topped anticline or double monocline, whose axis trends northwest-southeast. In Figure 4 this anticline is drawn with a large vertical exaggeration. The southern limb strikes N. 20° W. and dips 10° SW. The northern limb strikes N. 65° W. and dips 6° NE. The existence of this anticline had previously been recognized by Dr. Parry Reiche and Dr. Frank Johnson.

The anticline raises the base of the Mesaverde high above stream grade and exposes the Mancos shale in the valley walls. The massive white sandstone at the top of the lower part of the Mesaverde supports a prominent wide rock bench flanking the inner valleys. The same anticline may be traced southeastward to Balakai Mesa where it is truncated by the Hopi Buttes surface.

*Erosion surface.*—Measurement of altitudes at intervals at the top of the Oraibi-Wepo divide along the traverse A-B (Fig. 2) reveals the extent to which the structure has affected the topography. The topography of the northern two-thirds of this mesa is controlled by rock structure, and the profile of the stream divide parallels the rock strata (Fig. 4). South of Zihl-dush-jhini Peak, however, the strata are truncated, and this region probably represents an erosion surface which may be the Hopi Buttes surface or which may be a younger surface below it. At the end of Second Mesa, near the village of Shungopovi, as on Antelope Mesa, only the lower part of the Mesaverde remains. This may represent a stripped surface corresponding to the surface on Antelope Mesa to the east on which are found patches of the Bidahochi formation.

Thus during the Pliocene the southern edge of Black Mesa was an erosion surface and may have become buried by the Bidahochi formation. The greater portion of the mesa, however, was not eroded to a surface of low relief and was probably never buried but rose above the swampy lowlands to the south.

### HOPI BUTTES SURFACE IN THE HOPI BUTTES

*General statement.*—The marked unconformity between the Bidahochi formation and the underlying Mesozoic rocks is visible south of Black Mesa in the Hopi Buttes area, and in most places shows no evidence of relief during the Pliocene. At several localities, however, the Bidahochi formation evidently buried low escarpments held up by resistant Mesozoic strata.

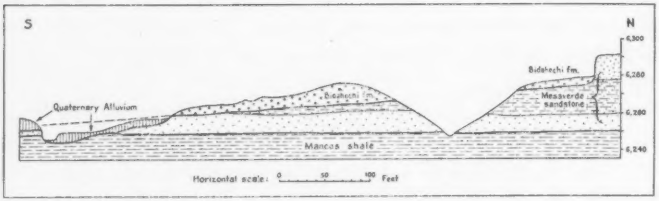


FIGURE 5.—*Cross section at locality C (Pl. 1)*
Showing relief on Hopi Buttes erosion surface.

*Escarpments formed by Mesaverde sandstone.*—At locality C (Pl. 1) the Hopi formation overlaps the Mesaverde sandstone, and where this overlap is exposed in a stream valley the Mesaverde forms a buried cliff about 10 feet high and in a distance of about 600 feet the total relief was about 40 feet (Fig. 5). The 10-foot cliff thus rose above a gentle shale slope. The sandstone at this locality forms the base of the Mesaverde and caps the mesa north of locality C (Fig. 2). North of the Jeddito Valley the whole lower part of the Mesaverde sandstone, 250 feet thick, is preserved. Thus another escarpment or series of escarpments once existed in between, at the site of the present Jeddito Valley.

In the headwaters of the Beshbito Wash, at locality D (Fig. 2), the basal sandstone of the Mesaverde forms a buried cliff over 50 feet high, as shown in Figure 6. This section is well exposed in a canyon wall and is probably the same escarpment as the one shown at locality C, farther west.

*Escarpment formed by Dakota sandstone.*—The resistant Dakota sandstone may also have formed a low escarpment before the Bidahochi formation was deposited. The outcrop of the Dakota sandstone occupies a large area near Steamboat Trading Post (*see* Gregory, 1917, Pl. 2) where the Bidahochi formation has been eroded off. It is evident that a wide rock bench underlain by the Dakota sandstone once existed. Such benches usually cap low escarpments and are associated with a steplike topog-

raphy. Thus in the Pliocene the Dakota sandstone probably formed a low escarpment.

### PLIOCENE TOPOGRAPHY

During the Pliocene, before the Bidahochi formation was deposited, the region of Black Mesa and the Hopi Buttes area had little relief, but there were low escarpments, some 50 feet high which encircled the Tusa-

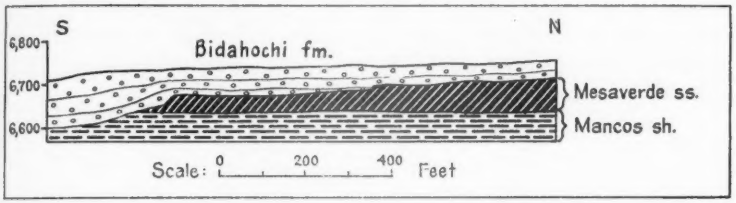


FIGURE 6.—*Cross section at locality D (Fig. 2)*
Showing relief of Hopi Buttes erosion surface.

yan downwarp as do the escarpments today. This contrasts greatly, however, with the relief of the present land surface, for valleys with walls 400 to 1000 feet high are characteristic.

Black Mesa in the early Pliocene was a highland area overlooking the lower country to the south. During this period aggradation by streams occurred and the southern edge of Black Mesa and the Hopi Buttes area were buried by sediments. Most of Black Mesa rose above this aggrading plain, however, and probably furnished material to be deposited on it.

In either Pliocene or post-Pliocene time, the Hopi Buttes surface and the overlying Hopi formation were deformed, and the gradients of the southwestward flowing streams were steepened. This warping may have accompanied the volcanic activity of the Hopi Buttes area.

### CORRELATION WITH OTHER EROSION SURFACES

In the Zuni area of west-central New Mexico, McCann (1938) showed that Tertiary sedimentary rocks believed to be Pliocene overlie a deformed erosion surface of low relief called the Zuni erosion surface. His argument for deformation is the same as that presented here for the deformation of the Bidahochi formation, that is, that the material which rests on the Zuni erosion surface is too fine-grained to have been deposited on the existing gradient. The Tertiary rocks of the Zuni area are apparently correlative with the Bidahochi formation in the Hopi Buttes for the same rocks crop out over a wide area between the two regions. Thus

the Zuni erosion surface is evidently correlative with the Hopi Buttes surface.

The late Tertiary lavas of the San Francisco volcanic field and the Coconino Plateau are believed by Robinson (1907) to rest on a Pliocene erosion surface. This may be correlative with the Hopi Buttes and Zuni erosion surfaces. The Chuska sandstone of the Chuska Mountains rests on an erosion surface of low relief. According to Reiche (1941) the Chuska sandstone is similar to the sediments which rest on the Zuni erosion surface, and is probably correlative with them. It is therefore probable that the erosion surface beneath the Chuska sandstone is correlative with the Hopi Buttes surface.

## VOLCANISM

### GENERAL CONSIDERATIONS

The diatremes or explosion pipes of the Hopi Buttes occur in a dense cluster covering about 800 square miles. A few are far removed from this cluster, as in the Jeddito and Oraibi valleys. The larger part of the main volcanic area has been mapped and the various diatremes located (Pl. 1). Detailed studies were made of several of the best exposed diatremes, which are numbered on the geologic map.

The central and northern portions of the Hopi Buttes have probably been relatively little dissected since completion of the volcanic activity and deposition of the Bidahochi formation. This area consists of a dense group of diatremes, many of which have poured forth dome-shaped or mushroom-shaped masses of lava. These have coalesced to give the appearance of a continuous sheet. To the south and west, where erosion has been more effective, only the feeding pipes of the diatremes are exposed, except here and there where a large lava dome or cluster of domes holds up a high mesa or butte. In both areas the nature of the filling of the volcanic vents ranges from basalt to the finest pyroclastic material. Extremely coarse explosion breccia occurs only in the deeply eroded diatremes.

In general the diameter of the vents decreases as erosion increases. In areas where dissection has been slight, at levels above the Hopi Buttes surface, explosion pits are generally 3000 to 4000 feet in diameter. In many places the initial explosion pit is overlain by domes of lava which have pushed outward, spilling over the sides, crumpling and pushing out the underlying and bordering tuffs and sediments. In a few places these lava eruptions were of sufficient duration to form rather continuous flows. The more deeply dissected diatremes range in diameter from 500 feet to 2000 or 3000 feet. In general the material in them is less well bedded, and, if pyroclastic, is coarser.

FIGURE 1. VIEW DOWN THE VALLEY OF COTTONWOOD WASH
From Bidahochi Butte, showing deeply eroded area.

FIGURE 2. BIDAHOCHI BUTTE VIEWED FROM INDIAN WELLS

FIGURE 3. STRATIFIED BASALTIC TUFF
Overlain by a lava flow on north side of Flat Mesa.

BUTTES AND STRATIFIED BASALTIC TUFF

FIGURE 1. SLICKENED SURFACE OF UPTURNED TRAVERTINE
At eastern margin of diatreme No. 6.

FIGURE 2. AGGLOMERATE AND LARGE BLOCKS OF NAVAJO SANDSTONE
At the contact of a diatreme with the Navajo sandstone.

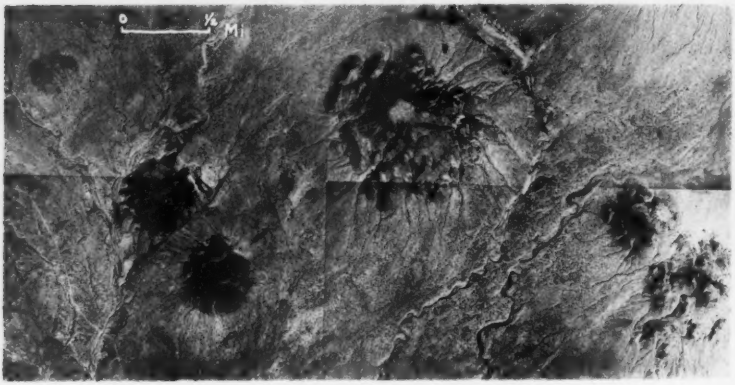


FIGURE 3. AIRPLANE PHOTOGRAPH OF DEEPLY ERODED VOLCANIC NECKS (Basalt-filled) near Dilcon (U. S. Soil Conservation Service).

UPTURNED TRAVERTINE, AGGLOMERATE, AND VOLCANIC NECKS

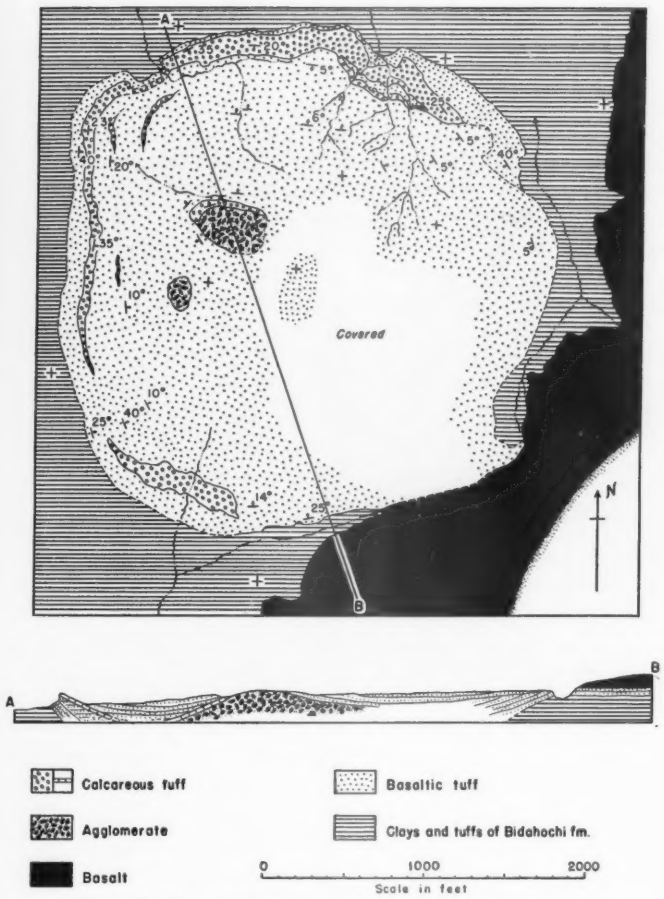


FIGURE 7.—*Map and structure section of diatreme number 1*
(Plane table survey, 1940.)

DIATREMES

NUMBER 1 (FIG. 7): This diatreme is one of the best exposed in the Hopi Buttes. It occurs in the north central portion of the Black Rock Buttes in an area underlain by water-laid, flat-lying basaltic tuff of the Bidahochi formation. The hard bed of tuff, altered to calcium carbonate which occurs near the periphery of the diatreme, is resistant and forms a nearly circular ridge, about 3000 feet in diameter; it can be clearly

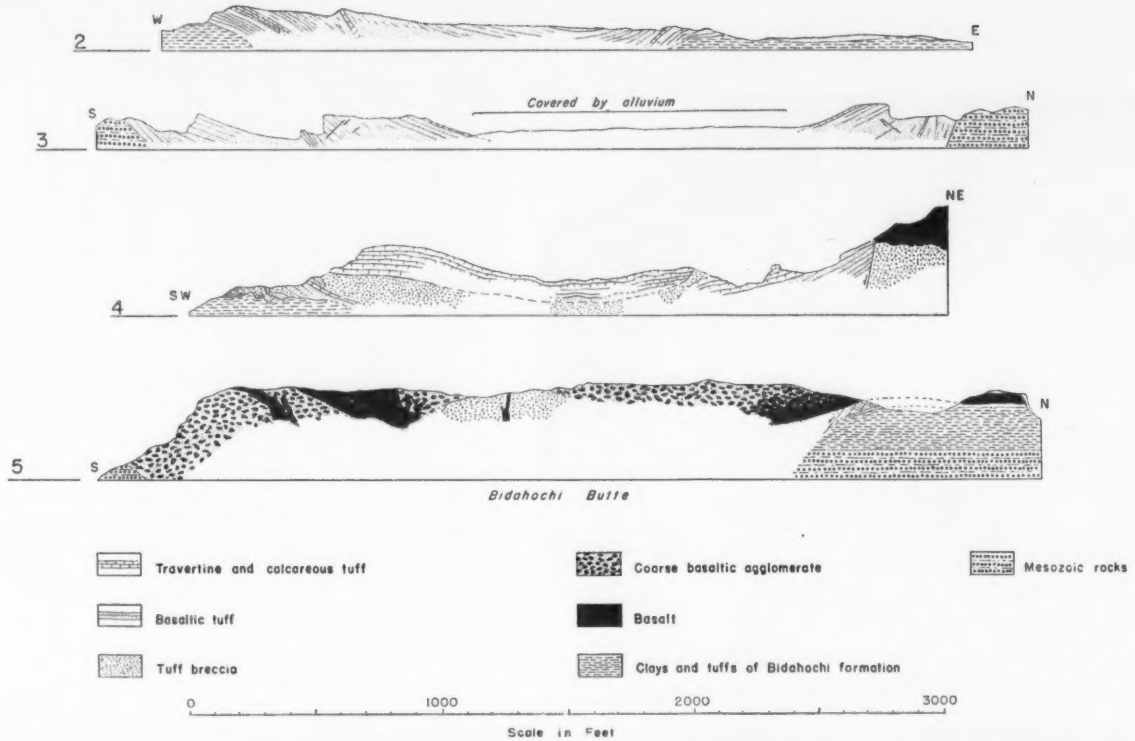


FIGURE 8.—*Cross sections of four diatremes*
Above or near level of Hopi Buttes erosion surface (Traverse with Paulin altimeter, 1940).

seen on an aerial photograph (Pl. 4, Fig. 1). The filling is mostly basaltic tuff ranging from coarse sand to fine ash. The only exception is the bed of altered white tuff which forms the ridge mentioned above.

The central outcrops of agglomerate are amygdaloidal limburgite with phenocrysts of augite. This rock is so vesicular that in this section the limburgite forms only a thin network around the cavities, which are filled with calcite. The white calcareous tuff at the periphery resembles travertine and is composed almost entirely of calcite with a grain size of less than .005 mm. A few coarse grains of quartz, augite, and limburgite are scattered throughout. The basaltic tuff forming much of the filling is indistinguishable from the tuff of the surrounding Bidahochi formation, except that it contains many layers of interbedded fine tuff, not as common in the rocks outside the vent. All the tuff is well stratified and appears to be water-laid. The grains consist of monchiquite, vesicular limburgite, augite, and quartz.

The structure is complex. At several places on the rim the outer wall of the diatreme can be observed. As indicated in the structure section the wall flares widely outward, dipping at an angle of from 5° to 40° toward the center. Inward from the outer wall several pronounced unconformities are observed in the tuffs. They suggest a complex history of formation which involves more than one explosion, or inward sliding of the filling. Toward the center, and around the inner core of basalt the tuffs dip outward almost as steeply as they dip inward at the periphery. A bed of fine, white calcareous tuff or travertine rests on the central outcrop of agglomerate. It may be the same bed as the similar tuff near the margin of the vent.

The structure of the diatreme must be explained by a complex history involving inward collapse of debris along the walls of the explosion pipe, later welling up of lava, and deposition of water-laid tuff, and partial subsidence. Part of the filling, especially the fine-grained material, may have been deposited while the small crater contained a shallow lake, and was supplied from outside the diatreme.

NUMBER 2 (FIG. 8): This diatreme lies a short distance southeast of number 1 and is about the same size. Like number 1 the Bidahochi sediments surround it. Its walls flare widely. The filling is mostly well stratified basaltic tuff. One bed of calcium carbonate which resembles travertine forms a resistant circular ridge. The strata dip inward, most steeply about 500 feet from the rim with dips as high as 65°, indicating subsidence of the filling. To the east of the diatreme, tuff beds dip outward, away from it. These may represent part of the rim of the original explosion crater.

FIGURE 8.—*Cross sections of four diatremes*
Above or near level of Hopi Buttes erosion surface (Traverse with Paulin altimeter, 1949).

NUMBER 3 (FIG. 8): This diatreme lies a few miles west of Indian Wells in the Lokasakad Valley. It is now exposed at a lower stratigraphic level than the diatremes described above and is larger than most of the diatremes exposed at this level. Williams (1936, p. 120) has described this diatreme as a cauldron subsidence. It contains over 1000 feet of basaltic tuff and interbedded calcareous tuff. The walls are very steep and a few hundred feet from the outer rim the tuff beds dip inward as much as 85° at one place. Lokasakad Wash has cut directly through the diatreme, cutting canyons in both sides of the rim, which disclose well-developed, but small faults, probably formed during the subsidence of the filling. The diatreme is surrounded by the Jurassic Navajo sandstone and probably is now exposed at least 100 feet below the level of its original crater.

NUMBER 4 (FIG. 8): Five miles north of Indian Wells on the east side of Bidahochi Wash is a diatreme which reveals more details of the manner of eruption. In the main portion of this well-exposed vent about 200 feet of fine, stratified, gray and buff-colored, calcareous tuff rests on coarse tuff-breccia which shows no evidence of bedding. At the western margin of the vent, coarse stratified basaltic tuff dips inward beneath the tuff-breccia. This is much contorted and rests on the soft clay of the Bidahochi formation. It apparently represents the tuff surrounding the original crater which was pushed outward or contorted by later eruptions, or by compression caused by the subsidence of the filling. At the northern margin the material of the diatreme rests against tuff-breccia and basalt of a high mesa. The contact is clearly a fault contact for the beds adjacent to the fault are curved upward, showing a downward movement of the material of the vent.

The basal tuff-breccia of the vent is interpreted as the filling of the crater, derived from an explosion in the diatreme itself. The fine-grained material above was probably washed or blown in from outside and may have been deposited in a crater lake. Subsidence of at least a portion of the diatreme occurred at some time during its formation. The original land surface must have been near the level of the basalt to the north, so that erosion has been considerable.

BIDAHOCHI BUTTE, NUMBER 5 (FIG. 8): Bidahochi Butte rises more than 600 feet above the surrounding low valleys. On first inspection this butte appears to be capped by a remnant of a sheet of lava. Its nearly circular plan, however, suggests a central core for the lava. Furthermore, as can be seen on airplane photographs, the system of gullies on top of the butte has a roughly circular outline indicating that the drainage has become adjusted to a basinlike rock structure. The detailed pace and compass traverse of the butte from south to north reveals a complex

diatreme. The fill is largely a coarse agglomerate composed of large, rounded, vesicular bombs. This material which shows no good stratification is penetrated here and there by dikes of coarse, olivine-rich basalt with orthoclase in the ground mass. At the northern end of the butte this material rests on stratified basaltic tuff which is faulted as though it had slid inward after becoming partially cemented. At the very northern edge of the butte some of the olivine basalt apparently spilled over the edge of the diatreme and is flat-lying or dips slightly outward. This butte is typical of many of the high, lava- and agglomerate-capped buttes of the region.

NUMBER 6 (FIG. 9): This diatreme lies a short distance north of Bidahochi Butte in the flat floor of the low Bidahochi Valley. Its form is regularly elliptical and is unusually well exposed. Its walls dip more steeply on the northeast rim than on the southwest rim. The travertine which rims the diatreme on east and north stands up as a prominent, resistant ridge. In several places it shows a slickened surface on its outer wall indicating subsidence of the filling (Pl. 3, Fig. 1). The contact with the Navajo sandstone is clean cut, and the sandstone is undisturbed, except that it is slightly altered for a distance of about half an inch away from the travertine. The travertine appears to have replaced the basaltic tuff which lies adjacent to it. The two materials merge into each other. The pure travertine is from 1 to 3 feet thick and contains abundant fragments of basalt and euhedral crystals of augite. Inward for a distance of several feet the proportion of basalt and augite grains increases. The pure basaltic tuff is conformable with the travertine, and is cemented by calcium carbonate.

The gypsum in the center of the diatreme is the highest deposit of the well-stratified filling. It is composed of swallow-tail crystals and coarse grains of gypsum. The strata are highly contorted. Beneath the gypsum is a considerable thickness of stratified gypsum interbedded with basaltic tuff. This material may have been deposited in a crater lake. Both the calcium carbonate of the travertine and the calcium sulfate of the gypsum may be in part hydrothermal in origin.

NUMBER 7 (FIG. 10): This diatreme forms a low, lava-capped butte on the north side of the Breezy Waters Valley. It is surrounded by the clays and tuffs of the Bidahochi formation and is probably exposed at the horizon at which eruption occurred. This butte shows part of the outward-dipping rim of the original crater on its southeast margin. This outer rim was apparently destroyed by a second explosion forming an unconformity. The second explosion was succeeded by deposition of stratified basaltic tuff, and extrusion of a thin cap of lava. Eruption of lava

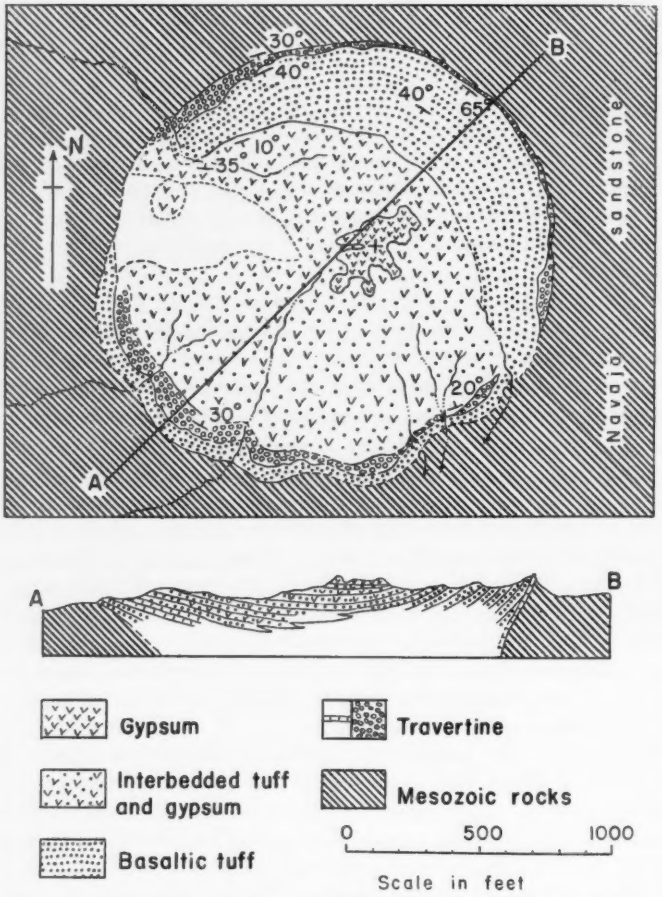


FIGURE 9.—*Map and cross section of diatreme number 6, Bidahochi Valley*
(Plane table survey, 1940.)

was followed by more explosions, forming two smaller vents. The eastern vent may have erupted the lava.

NUMBER 8 (FIG. 11): This is a small, deeply eroded vent at Indian Wells, surrounded by the Navajo sandstone. The exposed southern wall is vertical. The first eruption was explosive, choking the vent at this depth with coarse tuff-breccia and agglomerate. It was followed by a quieter eruption of monchiquite.

NUMBER 9 (FIG. 11): This diatreme lies about 500 feet northwest of number 8. As in number 8, the first eruption was explosive. It was followed by the eruption of coarse agglomerate which shows well-developed nearly vertical flow layers. Note that the basaltic tuff which forms the eastern rim dips steeply inward.

NUMBER 10 (FIG. 11): South of Black Rock Buttes on the north side of Lokasakad Valley is a diatreme which shows unmistakable evidence of subsidence. Only the northern half is exposed, the southern portion being eroded and buried by alluvium. A dike of vesicular, glassy limburgite extends northwestward from the diatreme.

The lowest beds of the filling consist of coarse breccia composed of blocks of stratified fine-grained tuff, basaltic tuff containing many boulders of glassy limburgite, and red Navajo sandstone. This coarse material probably represents the material which fell into the vent after the initial explosion. The coarse breccia is overlain by well-stratified and fissile, fine-grained, white, calcareous tuff. This in turn is overlain by inward-dipping, interbedded basaltic and fine-grained tuff. The coarser layers contain lenses of gravel with angular pebbles of Navajo sandstone and limburgite. The inward-dipping material perhaps fell in on top of the more fine-grained material during or after the subsidence. A fact of importance is that the boulders in the diatreme are composed of glassy limburgite similar to the dike which juts to the northwest from the diatreme. The lava flow on top of the mesa to the north is composed of coarse monchiquite and probably came from another vent.

The fault surface along the northern periphery is well exposed, showing a smooth, slickened surface. The lowest part of the fill has clearly subsided into the original vent, not by progressive slumping, but as a single unit.

NUMBER 11 (FIG. 11): This diatreme lies near Castle Butte Trading Post (Pl. 5, Fig. 2). Extending southwestward from it is a dike of dense agglomerate. The diatreme is exposed several hundred feet below the Hopi Buttes erosion surface and is thus deeply eroded. Like number 10 it contains both coarse explosion breccia and fine-grained calcareous tuff. There is no evidence of subsidence.

NUMBER 12 (FIG. 11): This deeply eroded diatreme lies near Pumpkin Flower Butte at Cottonwood Spring. Most of the fill is coarse, scoriaceous agglomerate which contains clearly exposed flow layers, showing evidence of subsidence at the periphery. At the south and west margins the agglomerate is underlain by stratified explosion breccia containing a high percentage of boulders of scoria and Mesozoic sandstone.

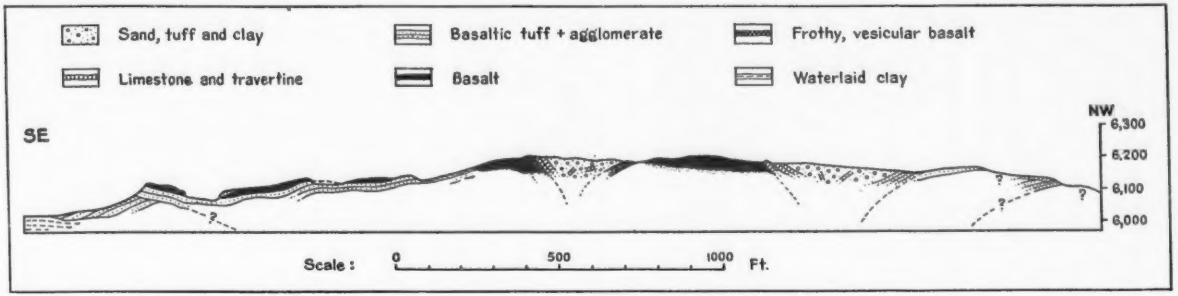


FIGURE 10.—*Cross section of diatreme number 7, Breezy Waters Valley*
(Traverse with Paulin altimeter, 1939.)

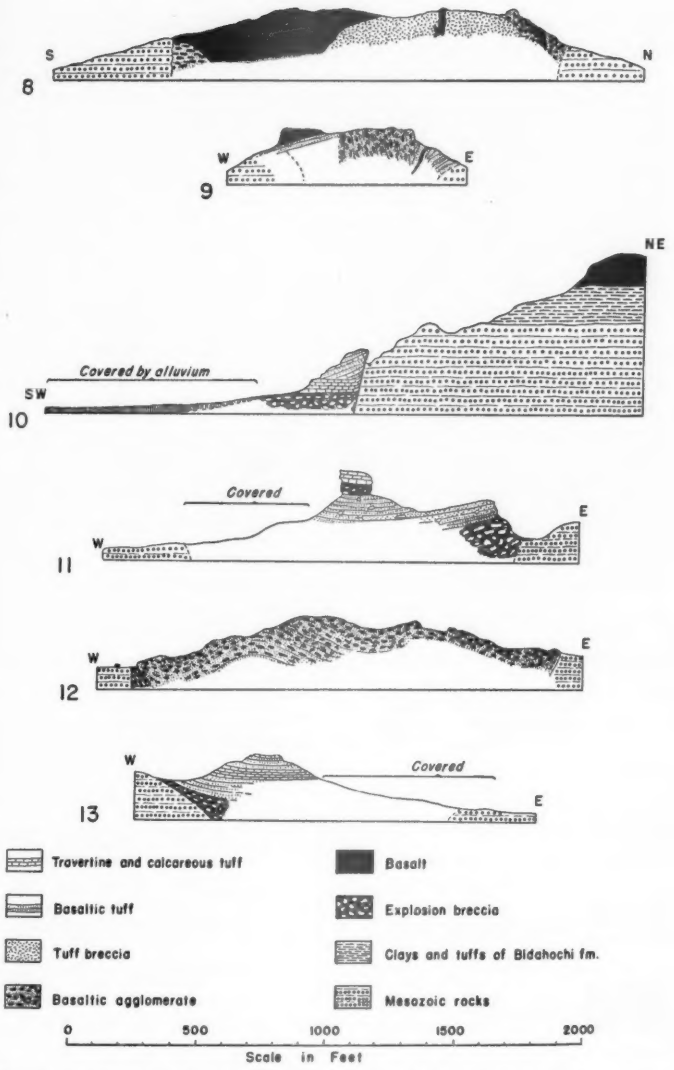


FIGURE 11.—*Cross sections of six diatremes eroded below Hopi Buttes surface*
(Traverses with Paulin altimeter.)

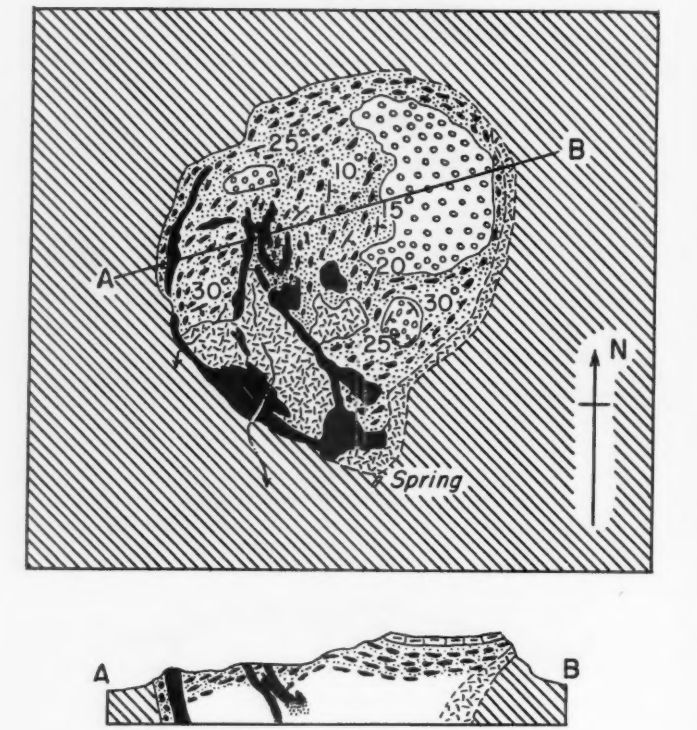


FIGURE 12.—*Map and cross section of diatreme number 14*
Near Castle Butte Trading Post (Plane table survey, 1940).

NUMBER 13 (FIG. 12): Near number 12 and at the same deep horizon is a small diatreme filled completely with clastic debris. Like number 11 it contains coarse explosion breccia as well as extremely fine-grained and stratified tuff.

NUMBER 14 (FIG. 12): Diatreme number 14 is eroded to a depth of 100 to 200 feet below the Hopi Buttes surface. It is a relatively small vest filled with a great variety of materials. The lower part of the filling consists of coarse agglomerate and tuff-breccia, penetrated by irregular dikes and stringers of monchiquite. The flow-banding of the agglomerate has a rough basin-shaped structure. The agglomerate is overlain by thin-bedded calcareous tuff, which is silicified at the top, and is hard. On the western periphery inward-dipping tuff-breccia and basaltic tuff underlie the agglomerate.

NUMBER 15 (FIG. 13 AND PL. 5, FIG. 1): This group or line of diatremes occurs at the southern margin of the area mapped. It forms the most deeply eroded group of diatremes observed, lying 700 or 800 feet below the Hopi Buttes surface. Three diatremes occur in a northwest-southeast line, two of them connected by a dike of agglomerate. The northwesterly and southeasterly vents are filled completely with structureless tuff-breccia. The middle one contains three materials; its center is composed of coarse tuff-breccia penetrated and welded by small apophyses of glassy limburgite. This material is resistant and stands up as a small castellated butte. The central core is surrounded by a less resistant fill of large blocks of Navajo sandstone, tuff, and tuff-breccia. The dike connecting the two vents is composed of limburgite with a large fragmental admixture of quartz and tuff. The central diatreme shows a simple history, an initial gaseous explosion, blowing out the explosion pipe, followed by more quiescent eruption of lava.

NUMBER 16 (NOT ILLUSTRATED): A well-exposed, tuff-filled diatreme occurs south of Cottonwood Wash, 3½ miles west of the Holbrook-Indian Wells highway. The borders of the vent are covered, but small canyons have exposed about 50 feet of the filling. In the center of the vent, the top bed is coarse-grained gypsum which is severely contorted, but more or less flat-lying. This is underlain by a grayish rock composed of fine-grained gypsum, calcareous tuff, and clay; a network of seams filled with gypsum so fine and clear that it might be called alabaster penetrates this rock. This is underlain by 20 feet of tuff-breccia, clay, fragments of Mesozoic sandstone, and fragments of cemented basaltic tuff. The tuff-breccia is permeated by gypsum which fills the pores between the grains. It appears that gypsum may have been deposited in a lake on top of the vent filling. Solutions percolating either upward or downward have deposited gypsum in pores and cracks in the underlying material.

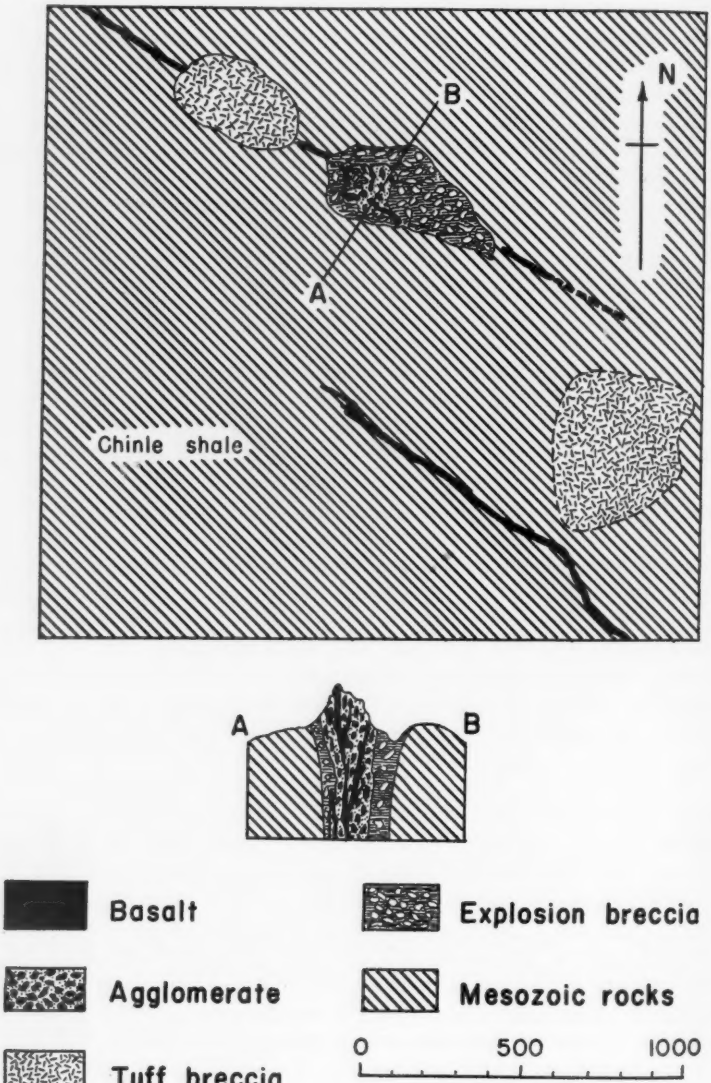


FIGURE 13.—*Map and cross section of three diatremes near Long Butte*
(Plane table survey, 1940.)

### DIKES

Dikes are more numerous in the areas where the Bidahochi formation has been eroded than at higher stratigraphic levels. However, they are surprisingly scarce anywhere. Most dikes are connected to at least one diatreme, and with one exception they trend northwest. At Castle Butte Trading Post is a dike which trends northeast.

The material which fills the dikes may be tuff-breccia, agglomerate, or basalt. A dike in the Breezy Waters Valley contains basaltic tuff.

### LAVA FLOWS

In general the lava erupted from the diatremes did not extend far from its source. Most of the eruptions formed flat, mushroom-shaped lava domes, which spilled only a short distance, if at all, over the crater rims. There are, however, a few lava flows. The largest of these forms a high mesa in the Black Rock Buttes, known as Flat Mesa. It has a maximum thickness of about 20 feet, and overlies nearly flat, stratified basaltic tuff of the Bidahochi formation. That this lava is a flow and not a series of lava domes is suggested by the fact that the mesa has an irregular outline, is flat on top, sloping with an eastward gradient of less than 2°, and its drainage pattern is dendritic (Pl. 5, Fig. 3). The lava was probably derived from a vent lying on the high point of the Black Rock Buttes to the west (Pl. 1, loc. E). There are a few other small areas apparently remnants of independent lava flows.

### COMPLEX STRUCTURE OF BLACK ROCK BUTTES

Northwest of Indian Wells, east of Cedar Springs and south of Breezy Waters Valley is a large highland area underlain mostly by basalt and agglomerate. This highland has been called Hauke Mesa by Gregory (1917). The Indian name for the area is "Tsejhin-deshgishih" which means Black Rock Buttes. This highland mass is underlain by the Bidahochi formation, which at this place contains a high proportion of lava and agglomerate. In nearly all places where erosion has cut canyons and valleys exposing the rocks beneath the highland, the structure is complex.

The most striking area is at Crazy Waters Spring, northwest of Indian Wells (Pl. 4, Fig. 2). This area is underlain by pyroclastic material and basalt. Excellent exposures reveal more than five diatremes. They are so close together that some overlap, and one truncates another. In one of them are several red graphic granite bombs more than a foot in diameter. The granite is composed of crystals of quartz, feldspar, and biotite which average over 2 mm. in diameter. Granite bombs of smaller size are reported by Williams (1936, p. 128) to occur at diatreme number 3 which lies nearby at Lokasakad Spring. No other localities have been

found. Williams believes that this granite must have been blown up from the pre-Cambrian basement, which in the Hopi Buttes probably lies at a depth of over 4000 feet. Perhaps the occurrence of these boulders in the area around Crazy Waters Spring is related to the great number of diatremes there as explosions may have been more violent and more frequent.

The group of diatremes at Crazy Waters Spring lies at the southern margin of the lava plateau of the Black Rock Buttes. This plateau has a rolling surface, consisting of rounded knobs of black, alkalic basalt or brownish agglomerate. Gullies in many places on the plateau show circular or radial patterns as on Bidahochi Butte. This suggests that they are subsequent streams adjusted to a complex underlying structure of closely spaced circular basins of tuff, agglomerate, and basalt or are consequent streams on lava domes. The only exception is the smooth lava flow of Flat Mesa. It is believed, therefore, that the Black Rock Buttes are a plateau consisting of closely spaced diatremes; the plateau has as yet been dissected only at the margin. Smaller basalt-capped mesas occur around the Black Rock Buttes, as for instance the mesa north of Pumpkin Flower Butte. These have a similar origin.

### SUMMARY OF DATA

The various diatremes apparently have many features in common and the pattern of the various eruptions was similar in every case.

(1) The diatremes not deeply eroded are in general larger in diameter than the deeply eroded ones. Furthermore, their walls flare widely, indicating that most of the diatremes are funnel-shaped and narrow downward. At depth their walls are perhaps more nearly vertical.

(2) The diatremes which are not deeply eroded are preponderantly filled with basaltic tuff, or with lava and agglomerate. Those which are deeply eroded are filled with lava, agglomerate, or very coarse explosion breccia and tuff-breccia showing little or no stratification. At depth, therefore, the diatremes contain mostly debris blown out by the initial explosion which has slumped into the opening, or they contain viscous erupted material which has displaced the explosion debris. At and near the surface the filling is stratified basaltic tuff, or fine calcareous tuff which has been blown or washed in from outside, or has been deposited in standing water.

(3) A few diatremes contain gypsum, travertine, or silica, minerals which may be of hydrothermal origin. Most diatremes contain tuff, which is altered to or cemented with calcium carbonate. These minerals must indicate that the vents contained circulating waters, which in part may have been juvenile.

(4) None of the diatremes disturbs the bedding of the intruded rocks.

(5) The material of many of the diatremes has subsided. In those diatremes which are not deeply eroded this is shown by steep inward dips of the filling, forming a structural basin (superbly illustrated by diatreme number 2). At greater depth the subsidence produced faults at the walls of the diatremes, and the filling moved downward more or less as a unit (as in diatremes 4 and 10). This difference in the manner of collapse is probably due to the flare of the diatremes near the surface. Material lying above the flare cannot move downward without slumping toward the center, and dragging along the walls.

(6) Most diatremes had a complex history. It is necessary to postulate more than one eruption to explain their structure.

(7) Many diatremes connect with dikes, or are with other diatremes aligned in long rows.

### SURFACE CONDITIONS DURING ERUPTION

*Origin of stratified tuff.*—The dominant material of the diatreme fillings near the surface is well-stratified tuff. This material is indistinguishable from the water-laid tuff which is widespread in the Bidahochi formation, interbedded with fluviatile sands and lacustrine clays. Stratified materials in explosion vents rarely occur. Bedding is not common at well-known explosion pipes such as Cripple Creek (Lindgren and Ransome, 1906), the Braden Crater, Chile (Lindgren and Bastin, 1922) or the diamond pipes of South Africa (Wagner, 1914). Walker (1928) in a paper summarizing the literature on explosion pipes makes no mention of bedding. Daly (1925, p. 26), however, has described a vent filled with inward-dipping stratified tuff on Ascension Island, and the Pretoria salt pan is a volcanic explosion pipe or diatreme filled with explosion debris interbedded with lake clays (Wagner, 1922). Daly and Wagner consider the bedding to be of fluviatile or lacustrine origin and the source of the material to be outside of the original crater. In the case of the Hopi diatremes the similarity between the tuffs in the vents, and the obviously water-laid ones of the Bidahochi formation, leads to the same conclusion.

*Origin of gypsum, travertine, and silica.*—Diatreme number 6 contains beds of gypsum and gypsiferous tuff in its center. At its eastern margin is a dike or vein of calcium carbonate which rests against the outer wall. This dike or vein is clearly secondary. Inward from the wall it becomes richer in basalt fragments, and at a distance of from 10 to 15 feet it is entirely basaltic tuff. This vein dips 65° and could not have been deposited in that position as a lacustrine marl. Apparently, therefore, it

is a vein of travertine intruded between the wall rock and the filling of the vent. It has altered part of the fill.

In diatreme number 14 the top layers of the filling are capped by chert which appears to be secondary, derived by the alteration of calcareous tuff to silica. The pure silica or chert is underlain by calcareous tuff which contains stringers and layers of silica and looks like a cherty limestone.

Gypsum occurs in the filling of diatreme 16. It forms cementing material in tuff-breccia and forms contorted beds of large, swallow-tail crystals.

The rocks of the Bidahochi formation are rich in calcium carbonate over a wide area, and the clay beds at its base contain gypsum crystals at many places. Chert, however, has not been found outside the diatremes.

The explanation for the occurrence of these three minerals is probably related to the high water table during the time of eruption. If the diatremes formed craters which held small lakes fed by ground water they must have become the site of deposition of compounds in the ground water. If the lakes had no outlets the water might have become supersaturated with these salts, depositing them on the crater floors. The diatremes themselves were certainly excellent channel-ways for percolating ground water, for many were filled with porous material. The ground water was hot and may have been added to by juvenile waters and by hot gases charged with calcium sulfate, or silica.

Large deposits of travertine, some with inward-dipping structure, occur in the Tertiary rocks south of the Navajo Country (Harrell and Eckel, 1939).

*Conclusion.*—During the period of eruption, the area of the Hopi Buttes was the center of an aggrading basin dotted with swamps and lakes. During times of flood, powerful streams carried large quantities of debris across the region, brought from highlands to the north and east. When eruptions occurred small craters were produced and large supplies of tuff were added to the aggrading streams. Inasmuch as the region had a high water table for at least parts of the year these craters probably at times contained standing water which allowed fine-grained tuff, calcium carbonate, and calcium sulfate to accumulate undisturbed. After a time aggradation and change of stream courses allowed the powerful streams to erode or spill over the crater rims and fill them with coarse material.

Many of the diatremes underlie flattish lava domes which welled outward, pushing the wet clayey material of the crater rim and the sur-

FIGURE 1. AIRPLANE PHOTOGRAPH OF DIATREME NO. 1
Showing basalt-capped mesa on right (U. S. Soil Conservation Service).

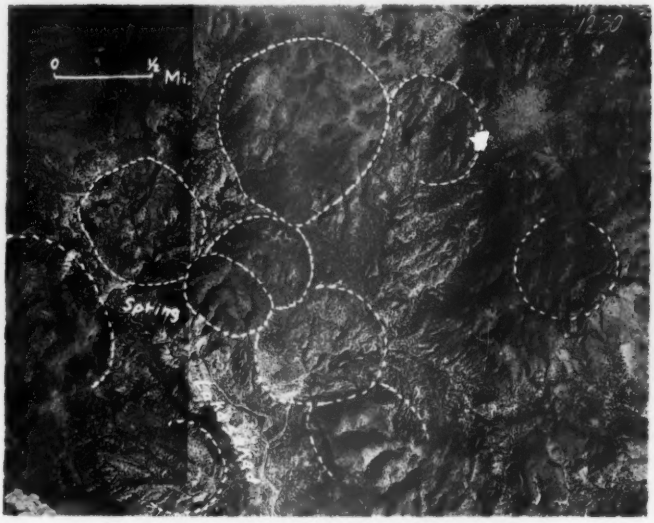


FIGURE 2. GROUP OF CLOSELY SPACED DIATREMES
On south margin of Black Rock Buttes at Crazy Waters Spring. The best exposed diatremes are outlined in white.

FIGURE 1. DIKE AND DIATREME AT DIATREME GROUP NO. 15

FIGURE 2. AGGLOMERATE-FILLED DIKE ON LEFT
Diatreme No. 11 in right center.

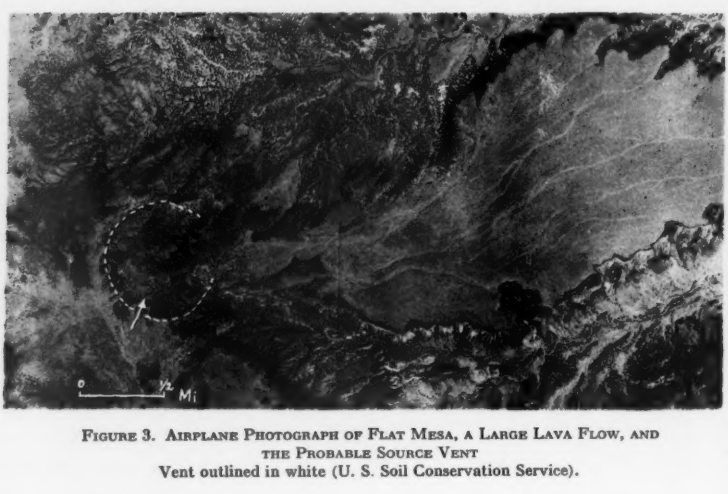


FIGURE 3. AIRPLANE PHOTOGRAPH OF FLAT MESA, A LARGE LAVA FLOW, AND THE PROBABLE SOURCE VENT
Vent outlined in white (U. S. Soil Conservation Service).

DIKES AND VIEW OF FLAT MESA

rounding plain ahead of them. Exposures of contorted clay and tuff at the borders of steep-faced lava domes are common.

### CONDITIONS OF ERUPTION AT DEPTH

*General statement.*—The material which makes up the fill of the Hopi diatremes at depth is not unusual. Other areas of diatremes show a similar structure and eruptive history, as for instance, the large group in Missouri (Rust, 1937), the Scottish diatremes (Geikie, 1897), and the numerous pipes described by Walker (1928). However, three features of the Hopi diatremes, which are probably not unique, are so well exposed that they are significant—the presence of connecting dikes and fissures, the pronounced funnel shape, and the evidence of subsidence of the filling.

*Alignment along fissures.*—In Figure 14 all the diatremes discovered are shown, whether they are exposed at the surface or are covered by basalt domes, and merely indicated by the presence of subsequent streams with circular courses. Many of them are arranged in long lines. Careful examination shows that a northwest-southeast alignment is most prominent. That this arrangement is not accidental is shown by the perfect alignment of several groups which are detached from the main volcanic area. Furthermore, some of the diatremes are attached to dikes filled either with clastic debris or with monchiquite and agglomerate, as in the case of diatreme group number 15 (Fig. 13). Thus, apparently, the diatremes are related to a system of fractures and dikes in the earth's crust. This relationship has been suggested by Dutton (1885) and Johnson (1907) for the volcanic necks of the Mount Taylor region, New Mexico.

*Funnel shape.*—Daubrée (1891), first to use the term "diatreme," produced artificial explosion pipes which were funnel-shaped at both ends. The Hopi diatremes appear to narrow downward and may even merge with a system of narrow fractures. Near the surface, however, they have a funnel shape approximating the funnel illustrated by Daubrée. Some of them flare widely at angles as low as 25° in the first 100 feet beneath the surface.

The origin of the flare may be related in part to the fact that near the surface the explosions penetrated soft, unconsolidated sediments, and in part to decrease in static pressure.

*Subsidence of filling.*—The subsidence, or inward collapse observed in most diatremes is observed in many volcanic areas. Many volcanoes have subsided by faulting after the original eruption. Obviously in the Hopi Buttes, subsidence occurred long after the first explosion, for in

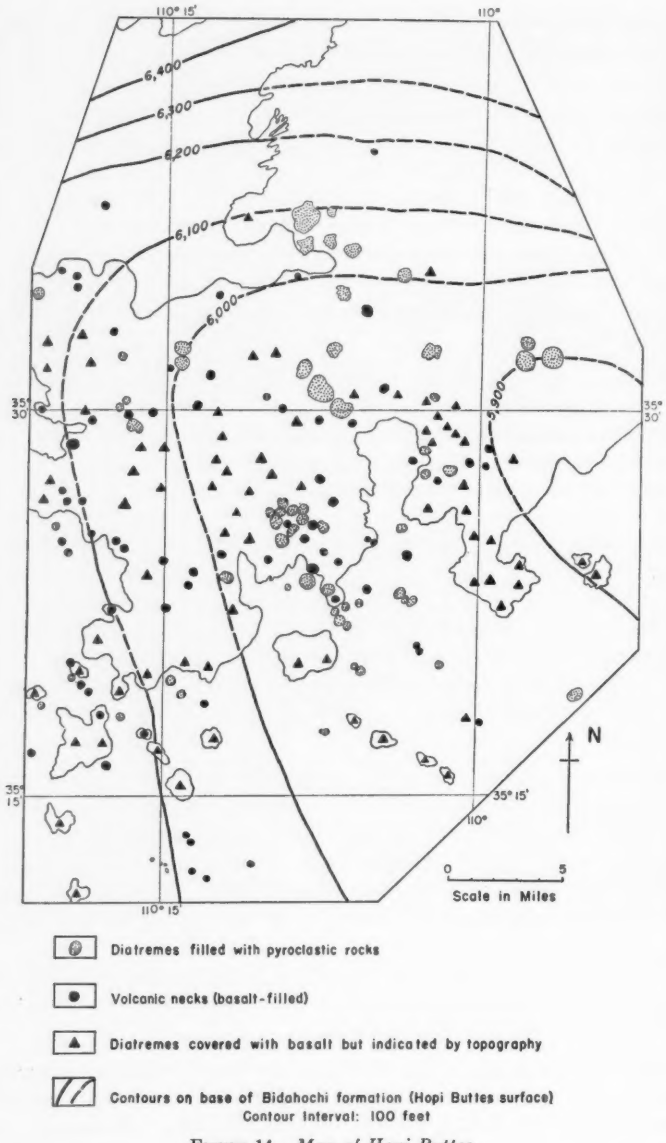


FIGURE 14.—*Map of Hopi Buttes*

Showing location of diatremes (tuff-filled vents drawn to scale) and contours on Hopi Buttes surface. Based on photoengravings of U. S. Soil Conservation Service.

most vents a considerable thickness of water-laid debris had accumulated. As shown by the flow layers of diatremes number 12 and 14, this subsidence occurred while the agglomerate filling was still plastic. In number 14, not only does the tuff have a basin structure, but also the agglomerate beneath it has a basin structure. This suggests that subsidence may have occurred because of the removal of molten material, or the contraction of molten material at depth.

Probably the filling of the diatremes was supported for a considerable time by a plug of cooling but hot and viscous material which, unless it was blown out by a subsequent eruption, subsided when the pressure of gas or molten magma was transferred to another diatreme or to another fracture.

### THEORETICAL CONSIDERATIONS

*Nature of volcanism.*—Williams (1936, p. 171) considers the eruptions in the Hopi Buttes to have occurred at relatively low temperature, and to have been of the phreatic type. The author agrees with this conclusion. Very little evidence of metamorphism around diatremes is observed in the Hopi Buttes. Diatreme 6, in the Bidahochi Valley is walled on the east side by a vein of travertine which has altered the Mesozoic sandstone adjacent to it, for a distance of about half an inch outward. This alteration is probably a result of hydrothermal activity, and deposition of calcium carbonate in the pores of the sandstone. Many xenoliths of sandstone are observed in other diatremes which appear to be completely unaltered.

During the period of volcanism the Hopi Buttes area was an aggrading basin of sedimentation, and was in part wet and swampy. The high content of calcium carbonate and gypsum in many of the vents is evidence that the water table was at times high, and that ground water circulated through the rocks beneath, using the diatremes as channelways. Conditions were therefore suitable for the generation of steam by contact with rising magma and the conclusion that the volcanism was phreatic is justified by *a priori* evidence.

*Magma reservoir.*—There is little positive evidence in the Hopi Buttes concerning the nature of the magma source. The field relationships indicate only that within a distance of about 1000 feet from the surface the diatremes narrow downward and are associated with a system of fractures in the earth's crust. The occurrence of numerous and closely spaced phreatic eruptions, however, is a phenomenon interpreted by many authors as associated with the occurrence of a laccolithic or sill-like magma reservoir at shallow depth. (For a concise discussion of this problem, and bibliography, *see* Daly, 1933.) The most famous example

is the large area of diatremes or embryonic volcanoes in Swabia (Branco, 1894).

The geomorphic history of the Hopi Buttes negates this hypothesis. It has been shown that before volcanism occurred, the Hopi Buttes area was an erosion surface of low relief. Contemporaneously with the volcanism it became an aggrading basin of sedimentation and the earliest sediments were deposited on a gradient less than the one existing today. During or after the volcanism the region was downwarped.

If the Hopi Buttes were underlain by a shallow, more or less horizontal and concordant magma reservoir, one would expect it to be domed, at least a perceptible amount. The opposite is true. A preferred conclusion is that the magma reservoir lies at greater depth and may even be the reservoir which furnished magma to the other Tertiary volcanic areas of the Navajo Country.

Several other areas in the Navajo Country contain Tertiary igneous rocks. Two of them, Navajo Mountain and Carrizo Mountain, are laccoliths, with no evidence of explosive volcanism. Others, although containing rocks of different composition, are similar to the Hopi Buttes area in that they contain diatremes which have not deformed or altered the rocks which they intrude. Rough calculation of the volume of the lava and other extruded debris of the Hopi Buttes indicates that it has a volume of about 50 to 100 cubic miles. Relatively little loss by dissection has occurred. The cross sections and map of Gregory (1917) indicate that both the Navajo Mountain and Carrizo Mountain laccoliths contain a similar volume of igneous rock.

In the Hopi Buttes the magma reached the surface by penetrating a widespread system of fractures. In the laccoliths the same volume of magma was confined to a domal reservoir at shallow depth.

## SUMMARY

In the Pliocene the Hopi Buttes area was reduced to an erosion surface of low relief, called the Hopi Buttes surface. This surface, whether or not it can be called a peneplain, must have been reduced nearly to base level. At present it has an average gradient of 50 feet per mile. Inasmuch as the surface has been deformed the Pliocene gradient must have been less. Low escarpments rose above the plain, however, and the central part of Black Mesa formed a residual highland mass.

Pliocene sediments were deposited in a broad lowland, extending from the Hopi Buttes to a point far to the southeast, probably to the Zuni Mountains. In the Hopi Buttes the first sediments were clays, deposited in ephemeral lakes and swamps. At a later time coarse materials were spread out by aggrading streams.

This period of sedimentation was accompanied by volcanic activity. Magma penetrated the earth's crust beneath the area in a complicated system of fissures. As the rising magma neared or reached the surface its extrusion was accompanied by violent explosions which produced numerous, funnel-shaped diatremes. The craters of the diatremes were partially filled soon after the explosions by the rising magma or by a mass of explosion debris and viscous agglomerate, penetrated by apophyses of basalt. Many of them had a crater rim at the surface and contained pools of water which allowed fine ash, gypsum, and other materials to collect. Eventually many of the craters which had not overflowed with lava or agglomerate were filled and buried by coarse sandy basaltic tuff, carried by aggrading streams.

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Hofstra College, Hempstead, N. Y.
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